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CONTENTS

Determination of the Vertical Eddy Viscosity in Sea Water from Wind Stresses and Surface Current Velocities	K. HIDAHA, T. MOMOI..	1
Calculation of Upwelling	K. HIDAHA..	11
A Contribution to the Computation of Three-Dimensional Ocean Currents by High-Speed Computers.....	K. HIDAHA..	16
Equatorial Flow and Inertia Terms	K. HIDAHA..	29
Sea Surface Temperature, Radiation, and Wind Anomalies in the North Pacific Ocean.....	G. I. RUDEN, J. L. REID..	36
Balance of the Transport of Sediment due to Wave Action in Shoaling Water, Surf Zone and on Foreshore.....	Y. NAGATA..	53
An Analysis of the Mechanism of the Cold and Warm Water Masses in the Seas adjacent to Japan	J. FUKUOKA..	63
Some Calculations on the Equatorial Circulation	K. YOSHIDA..	101
Determination of Uranium in Sea-Water Using a Polarographic Method M. ISHIBASHI, T. FUJINAGA, K. IZUTSU, T. YAMAMOTO, H. TAMURA..		106
The Relation between Chlorinity and Silicate Concentration of Waters observed in Some Estuaries V	H. MAEDA, K. TAKESUE..	112
Uses of Radio-isotopes for Tracing Predations of Trout-Fry..	Y. HIYAMA..	120
Appendix Contents of Previous Issues of the Records of Oceanographic Works in Japan.....		133



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Determination of the Vertical Eddy Viscosity in Sea Water from Wind Stresses and Surface Current Velocities

By

Koji HIDAOKA and Takao MOMOI

(Geophysical Institute, Tokyo University, Tokyo)

Abstract

A relation was derived, from EKMAN's theory of wind-driven ocean currents, between the components of surface current velocity, the coefficient of vertical eddy viscosity and wind stress exerted on the sea surface. By substitution of observed velocity of surface current and wind stress data, magnitudes of the coefficient of vertical eddy viscosity were derived. A short discussion was made on the result.

1. Introduction

Although the idea of vertical mixing has played an important role in dynamical oceanography every since V. W. EKMAN propounded his well-known theory of drift currents, very few determinations were made of the magnitudes of coefficient of vertical mixing. In his original paper (EKMAN, 1905), EKMAN suggests that this quantity can be computed by an indirect determination of EKMAN's depth of frictional influence in an enclosed sea by using Colding's relation between the surface slope, depth to the bottom and surface departure from the undisturbed level. (A. COLDING, 1881) Thus the eddy viscosity was given as proportional to the square of the wind velocity. Afterwards H. THORADE concluded that this quantity is proportional to the cube of wind velocity for a wind velocity below 6 m/s (THORADE 1914, 1928).

SVERDRUP and FJELDSTAD computed the coefficient of eddy viscosity from the result of a current measurement at a station about 22 meters deep on the North Siberian Shelf, by numerically integrating the steady state equations of ocean currents and inserting the observed wind stress and the current velocities at several different depths. (SVERDRUP, 1929; FJELDSTAD, 1929). According to their results, the coefficient of eddy viscosity increases from a value less than 70 at the bottom to about 385 at the sea surface. Later, SUDA (SUDA, 1936) determined the coefficient of eddy viscosity from the result of his current measurements in the adjacent seas of Japan and obtained values ranging from 750 to 75000 c. g. s. There are some other results derived from tidal ob-

servations, most of them giving a common result that eddy viscosity increases from the bottom upwards.

There has been, however, no attempt to determine the eddy viscosity over an extensive oceanic areas, and to discuss this quantity statistically.

The present research is the result of a determination of the distribution of eddy viscosity over the surface of the Pacific Ocean, derived from the wind stress data and the velocities of surface ocean currents. The wind stresses are available from HIDAOKA's Table (HIDAOKA, 1958), and the surface ocean currents can be known from U.S. Navy Pilot Charts or other current charts published by several authorities. The formula for computing eddy viscosity was derived from EKMAN's expressions for current velocity in terms of wind stress and eddy viscosity.

2. Theoretical Analysis

To make the mathematical analysis simpler, let us with NOMITSU (NOMITSU, 1942) assume that the coefficient of vertical mixing has different values μ_1 and μ_2 in x - and y -directions. The steady state equations for drift currents will then be

$$\left. \begin{aligned} \mu_1 \frac{d^2 u}{dz^2} + 2\omega\rho \sin \varphi \cdot v &= 0; \\ \mu_2 \frac{d^2 v}{dz^2} - 2\omega\rho \sin \varphi \cdot u &= 0 \end{aligned} \right\} \quad (1)$$

where x -, y - and z -axis are taken positive eastward, northward and vertically downward respectively, u and v denote the x - and y -component of current velocity, ρ the density of sea water, ω the angular velocity of the Earth, and φ the latitude, positive and negative in the northern and southern hemisphere respectively.

In addition to these equations, we have to add the surface conditions:

$$z=0: -\mu_1 \frac{du}{dz} = \tau_x; \quad -\mu_2 \frac{dv}{dz} = \tau_y \quad (2)$$

meaning the stress in the water balances the wind stress τ_x and τ_y on the sea surface.

The bottom conditions are

$$z=\infty: u=0; \quad v=0 \quad (3)$$

showing that there is no motion of water at very deep levels.

Making use of these four conditions (2) and (3), we have by solving two equations of (1) for u and v

$$\begin{aligned} u &= \frac{e^{-\sqrt{a}z}}{2\sqrt{a}} \left\{ \left(\frac{\tau_x}{\mu_1} + \frac{\tau_y}{\mu_2} \frac{a_1}{a} \right) \cos \sqrt{a}z + \left(-\frac{\tau_x}{\mu_1} + \frac{\tau_y}{\mu_2} \frac{a_1}{a} \right) \sin \sqrt{a}z \right\} \\ v &= \frac{-e^{-\sqrt{a}z}}{2\sqrt{a}} \frac{a}{a_1} \left\{ \left(\frac{\tau_x}{\mu_1} - \frac{\tau_y}{\mu_2} \frac{a_1}{a} \right) \cos \sqrt{a}z + \left(\frac{\tau_x}{\mu_1} + \frac{\tau_y}{\mu_2} \frac{a_1}{a} \right) \sin \sqrt{a}z \right\} \end{aligned} \quad (4)$$

where

$$\left. \begin{aligned} a_1 &= \pm \frac{\omega |\sin \varphi| \cdot \rho}{\mu_1} \\ a &= \frac{\omega |\sin \varphi| \cdot \rho}{\sqrt{\mu_1 \mu_2}} \end{aligned} \right\} \quad (5)$$

If we denote by u_0 and v_0 the components of surface velocity, we have by putting $z=0$ in equations (4)

$$\left. \begin{aligned} u_0 &= \frac{1}{2\sqrt{a}} \left\{ \frac{\tau_x}{\mu_1} + \frac{\tau_y}{\mu_2} \frac{a_1}{a} \right\} \\ v_0 &= \frac{-1}{2\sqrt{a}} \frac{a}{a_1} \left\{ \frac{\tau_x}{\mu_1} - \frac{\tau_y}{\mu_2} \frac{a_1}{a} \right\} \end{aligned} \right\} \quad (6)$$

Now a great deal of data for the velocities of the surface current u_0 and v_0 and wind stresses τ_x and τ_y are available. Substituting (5) into (6) we have the following equations with two unknowns:

$$\left. \begin{aligned} \frac{\tau_x}{\mu_1} \pm \frac{\tau_y}{(\mu_1 \mu_2)^{\frac{1}{2}}} &= 2u_0 \frac{(\omega |\sin \varphi| \cdot \rho)^{\frac{1}{2}}}{(\mu_1 \mu_2)^{\frac{1}{4}}} \\ \frac{\tau_x}{(\mu_1 \mu_2)^{\frac{1}{2}}} + \frac{\tau_y}{\mu_2} &= 2v_0 \frac{(\omega |\sin \varphi| \cdot \rho)^{\frac{1}{2}}}{(\mu_1 \mu_2)^{\frac{1}{4}}} \end{aligned} \right\} \quad (7)$$

Solving these equations for μ_1 and μ_2 , we have

$$\begin{aligned} \mu_1 &= \frac{1}{4\omega\rho|\sin\varphi|} \frac{\tau_x^2}{u_0^2} \frac{2}{\pm \left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right) + \sqrt{\left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right)^2 + 4 \frac{\tau_y}{\tau_x} \frac{u_0}{v_0}}} \\ &\quad \left\{ \pm \left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right) + \sqrt{\left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right)^2 + 4 \frac{\tau_y}{\tau_x} \frac{u_0}{v_0}} \pm \frac{\tau_y}{\tau_x} \right\}^2 \\ \mu_2 &= \frac{1}{4\omega\rho|\sin\varphi|} \frac{\tau_x^2}{u_0^2} \frac{\pm \left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right) + \sqrt{\left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right)^2 + 4 \frac{\tau_y}{\tau_x} \frac{u_0}{v_0}}}{2} \\ &\quad \left\{ \pm \left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right) + \sqrt{\left(\frac{\tau_y}{\tau_x} + \frac{u_0}{v_0} \right)^2 + 4 \frac{\tau_y}{\tau_x} \frac{u_0}{v_0}} \pm \frac{\tau_y}{\tau_x} \right\}^2 \end{aligned} \quad (8)$$

These expressions enable us to compute the magnitudes μ_1 and μ_2 directly. The wind stress data were adopted from the Tables by HIDAOKA (HIDAOKA, 1958) and the surface current data were taken from the U.S. Navy Pilot Charts.

3. Numerical Computation

From expressions (8) we can compute the regional distribution for magnitudes of the coefficients of vertical mixing μ_1 and μ_2 . The geometric mean

Table 1-A. The Regional Distribution of Magnitudes of the Coefficients

	115°E	120°	125°	130°	135°	140°	145°	150°	155°
50°N						$\mu_1 =$		12	72
						$\mu_2 =$		43	93
45°						$\mu =$		23	82
							16	110	150
							701	420	2400
40°							106	214	600
						1	9	28	56
						0.01	9	432	2130
35°						0.1	9	110	109
				186	6	4	54	26	6
				191	4	1	2961	82	19
30°				188	5	2.5	400	51	13
		12	18	120	322	123	7	12	3
		14	11	193	1280	111	710	63	9
25°		13	14	150	640	118	70	27	5
		250	5	9	9	14	21	0.25	326
		10	9	28	45	98	62	270	28
20°		50	6.7	15	22	37	36	8.2	95
		120	3	31	16	31	36	28	150
		11	6	25	120	29	120	110	139
15°		36	4	28	44	30	65	55	144
			200	8	6	12	12	150	15
			1820	200	920	560	286	25000	11
10°			600	40	74	80	58	1940	13
			1	4.9	31	100	16	4	7
			8.001	4.2	63	10	0.5	1	28
5°			2.73	4.24	44	32	2.83	2.5	14
			161	5.6	6.1	13	26	6.1	160
			111	5.01	5.3	102	12	63	12
0°			133	5.1	5.57	36	17	19	44

Table 1-B. The Regional Distribution of Magnitudes of the Coefficients

	120°E	125°	130°	135°	140°	145°	150°	155°
0°				$\mu_1 =$	6.3	1.3	21	3
				$\mu_2 =$	7.3	0.0	101	5000
5°				$\mu =$	6.62	0.16	46	125
							6.04	32
							6.2	421
10°							6.09	116
							9	10
							31	14
15°							17	12
							80	11
							1200	36
20°							310	20
							21	89
							0.8	23
25°							4.1	45
							42	360
							11	14
30°							21	71
							25	98
							20	23
35°							22	47
	112	1110	150	7.2	66.2	21	27	6.01
	31	120	230	92	70.8	45	101	2.1
40°	59	365	186	15	60.4	31	5.2	6.14
	11	1800	1620	33	0.1	11	0.1	2
	15	18	230	640	530	9	0.6	0.15
45°S	12	180	610	145	7.3	10.4	0.34	0.55

Determination of the Vertical Eddy Viscosity in Sea Water from
Wind Stresses and Surface Current Velocities

5

of Vertical Mixing μ_1 , μ_2 and μ in the North Pacific.

160°	165°	170°	175°	E180°W	175°	170°	165°	160°	155°W
			$\mu_1 =$		0.0	28	9	48	55°N
			$\mu_2 =$		0.0	0.1	9	84	
			$\mu =$		0.0	1.8	9	63	50°
155	200	192	100	121	5.8	150	220	2.8	
110	200	393	10	484	682	110	220	80	
128	200	325	32	260	63	128	220	15	45°
380	52	36	34	23	4.8	56	69	24	
11	173	142	63	29	19	284	101	1120	
65	95	71	46	26	9.8	126	84	118	40°
113	92	36	61	37	26	84	13710	40	
121	2200	16	815	120	26	2090	11	380	
117	450	24	224	66	26	420	1280	122	35°
8.3	0.8	57	17	41	9.8	23	6.2	5.4	
2.7	2.6	236	43	130	101	243	283	210	
4.9	1.45	116	27	73	99	75	42	10.6	30°
10	50	24.3	8.03	28	57	1140	49	15.7	
14	20	1.2	6.1	0.9	1.6	42	14	1	
12	10	5.6	7.02	5.3	9.5	230	26	3.5	25°
92	821	208	35	42	36	48.3	66	37	
7260	342	831	32	6	55	450	261	182	
820	530	415	33	2.2	44	148	133	82	20°
11.1	14.4	4.5	4.6	69	7.2	2.3	30	224	
0.3	1	23	45	619	0.7	2.9	6000	931	
1.8	1.4	10.1	14.4	206	2.2	2.2	410	460	15°
39	12	61	21	2652	6.2	13	118	294	
12	62	0.7	43	43	923	223	1231	1021	
5.2	27	6.5	30	330	75	54	380	550	10°
241	300	960	452	4	13	325	320	850	
0.1	360	42	210	36	15	28	20	850	
5.4	316	200	310	13	14	96	80	850	5°
12	126	41	21	6.7	18.2	92	62	27	
62	6.2	151	92	356	82	730	470	283	
28	27	79	43	49	12	260	170	88	0°

of Vertical Mixing μ_1 , μ_2 and μ in the South Pacific.

160°	165°	170°	175°	E180°W	175°	170°	165°	160°	155°W
31	6.2	0.9	1.6	42	1.2	32	6.1	7.2	0°S
91	6.8	91	70	11	10	120	63	92	
53	6.4	9	16	22	3.4	61	19	15	5°
160	415	7	32	5	9	4.9	21	23	
430	415	710	11	7.2	6	45	45	213	
260	415	70	18	6	7.3	14	31	65	10°
28	326	8.6	27	42	10	15	7	36	
432	28	8.1	101	40	14	92	28	55	
110	95	8.2	5.2	41	12	37	14	44	15°
25	11	12	150	9	56	300	50	50	
456	36	43	2400	9	210	56000	210	120	
107	20	23	600	9	106	4100	102	77	20°
18	12	250	14	31	28	29	200	28	
11	14	10	98	52	432	1.3	285	45	
14	13	50	37	40	110	6.1	240	35	25°
0.2	326	21	42	89	28	200	31	1	
270	28	0.8	11	23	110	1820	55	436	
8.2	95	4.1	21	45	55	600	41	210	30°
19	12	11	0.8	11	11	13	4	45	
14	0.94	13	6.2	9	14	15	36	21	
15	3.9	12	2.2	10.4	12	14	13	31	35°
2.2	39	12	5	360	120	21	31	100	
0.01	12	62	12	14	11	45	63	10	
0.01	5.2	27	7.7	71	36	14	44	32	40°
81	14	9	4	322	2.8	250	16	28	
51	4.2	4	1	12	80	10	120	110	
64	7.7	6	2.5	64	15	50	44	55	45°S

μ_1
 μ_2
 μ

$\mu = \sqrt{\mu_1 \mu_2}$ will be taken as the probable magnitudes of the coefficient of vertical mixing. This quantity is computed for the Pacific area west of 155°W meridian, and is shown in the Tables 1-A and 1-B.

In a similar manner, we computed, in order to get an idea for the cases of pure drift currents, the regional distribution of magnitudes of the coefficients of vertical mixing μ_1 , μ_2 and μ corrected against geostrophic flows in the Pacific to the west of 155°W, and south of 20°N and north of 15°S. (Table 2) In these cases, geostrophic components were subtracted from the u_0 and v_0 given in the U.S. Navy Pilot Charts.

4. Discussion

Table 1 and 2 show us the regional distribution of the magnitudes of the coefficients of vertical mixing μ , both uncorrected and corrected against the geostrophic currents.

Table 3 shows frequency versus the magnitudes of μ for the N. and S. Pacific west of 155°W and the average for both hemispheres respectively. They are also represented in Fig. 1.

Table 3. Frequency versus the Magnitudes of μ Computed from
Observed Wind Stresses and Surface Currents.

μ \ frequency	N. Pacific	S. Pacific	Average
0—10 c. g. s.	26	24	25
10—20	7	16	10.5
20—30	8	4	6
30—40	5	7	6
40—50	6	8	7
50—60	3	4	3.5
60—70	5	2	3.5
70—80	6	2	4
80—90	5	0	2.5
90—100	4	0	2
100—110	3	2	2.5

Table 3-A. With a Smaller Interval of μ .

μ \ frequency	N. Pacific	S. Pacific	Average
0—2 c. g. s.	6	2	4
2—4	10	3	6.5
4—6	10	7	8.5
6—8	4	10	7
8—10	4	4	4

Table 4. Frequency versus the Magnitudes of μ Computed from Observed Wind Stresses and Surface Currents near the Equator.

μ	frequency	μ	frequency
0—10 c. g. s	17	100—110 c.g.s.	0
10—20	11	110—120	0
20—30	4	120—130	1
30—40	4	130—140	0
40—50	6	140—150	0
50—60	6	150—160	0
60—70	2	160—170	0
70—80	4	170—180	0
80—90	1	180—190	0
90—100	1	190—200	0

Table 5. Frequency versus the Magnitudes of μ Computed from Observed Wind stresses and Surface Currents corrected against Geostrophic Flows near the Equator.

μ	frequency	μ	frequency
0—10 c. g. s.	5	100—110 c. g. s.	3
10—20	8	110—120	4
20—30	5	120—130	2
30—40	3	130—140	2
40—50	3	140—150	1
50—60	2	150—160	2
60—70	3	160—170	1
70—80	3	170—180	2
80—90	2	180—190	1
90—100	2	190—200	0

When they are corrected against geostrophic flows, their maximum occurrence takes place in the range $\mu=4-6$ c. g. s. in the average for both hemispheres, while the maximum of μ exists within the ranges $\mu=2-6$ c. g. s. and $\mu=6-8$ in the North and South Pacific respectively.

Tables 4 and 5 show the curve—frequency versus the magnitudes of in the region west of 155°W , and bounded by parallels of 20°N and 15°S in the Pacific. The maximum magnitude of μ occurs for $\mu=0-10$ c. g. s. for currents not corrected against geostrophic flows, while maximum magnitude of μ , in the range $\mu=10-20$ c. g. s. for these corrected against geostrophic flows. (See Fig. 2)

5. Conclusion

By using the equations of drift currents (1) and the current data not corrected for geostrophic flows, the magnitudes of μ in the Pacific west of 155°W mostly exists in the range $\mu=4-6$ c. g. s.

When the current data are corrected for the geostrophic flows, the magnitude of μ exists in the range $\mu=10-20$ c. g. s. in the Pacific west of 155°W , bounded by the parallels of 20°N , and 15°S .

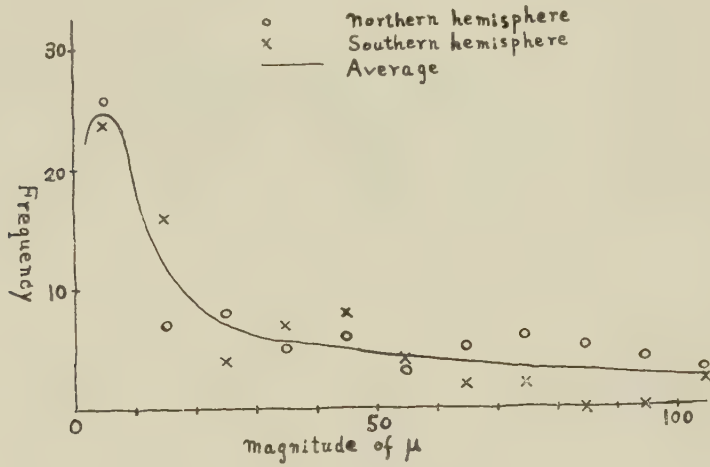


Fig. 1. Frequency versus Magnitudes of μ .

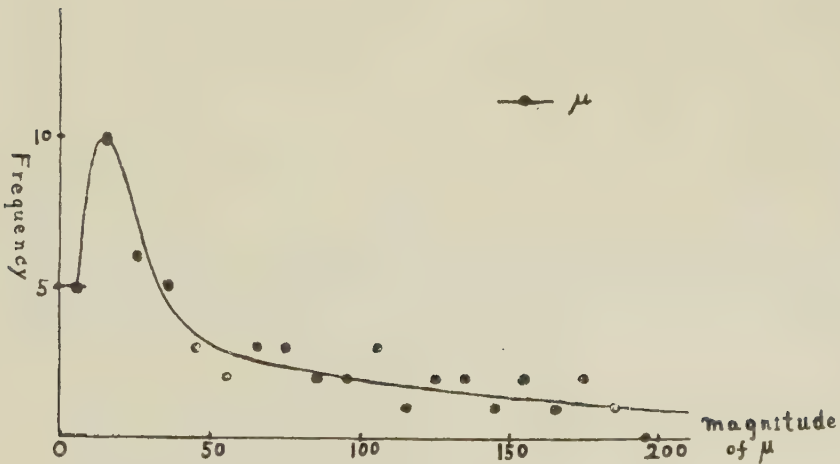


Fig. 2. Frequency versus Magnitudes of μ Computed from the Currents Corrected for Geostrophic Flows.

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Calculation of Upwelling

By

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Abstract

An expression to compute upwelling, valid except at the equator and poles was derived from the hydrodynamical equations, taking the effects of pressure gradients, CORIOLIS forces and wind stresses exerted on the surface of the sea. Several examples were worked out, use being made of the POFI-data.

Introduction

In this paper is given a simple method of computing upwelling and sinking, or the vertical velocity in a stationary motion of sea water, use being made of the hydrodynamical equations and taking the pressure gradients, the vertical mixing and Coriolis forces into account. The effects of thermodynamic processes, horizontal mixing and field accelerations are neglected.

A simple formula will be derived for computing an approximate value of vertical component of velocity in the sea, or the upwelling. This will fail at and around the equator and poles. Furthermore, it will be in some errors at and close to the sea surface and the bottom. Approximate estimate for these errors, however, will be possible.

The resulting formula looks similar to STOMMEL's, though the method of derivation is different. (STOMMEL 1956)

Taking the x -axis eastwards and y -axis northwards and z -axis positive downwards, the hydrodynamical equations of stationary ocean currents in the northern hemisphere are, when there is a certain pressure distribution in the sea whose surface is swept by prevailing winds:

$$\frac{\partial \tau_x}{\partial z} + 2\omega \rho \sin \varphi v = \frac{\partial p}{\partial x} \quad (1)$$

$$\frac{\partial \tau_y}{\partial z} - 2\omega \rho \sin \varphi u = \frac{\partial p}{\partial y} \quad (2)$$

where u and v are the components of current velocity parallel to x and y -axis, p the pressure, τ_x and τ_y are the components of stresses, ρ the density of sea

water, ω the angular velocity of the earth and φ the geographical latitude.

The equation of continuity is

$$\frac{\partial}{\partial x}(\rho u) + \frac{1}{\cos \varphi} \frac{\partial}{\partial y}(\rho v \cos \varphi) + \frac{\partial}{\partial z}(\rho w) = 0$$

or

$$\frac{\partial}{\partial x}(\rho u) + \frac{\partial}{\partial y}(\rho v) - \frac{\tan \varphi}{R} \rho v + \frac{\partial}{\partial z}(\rho w) = 0 \quad (3)$$

where w is the vertical component of water flow taken downwards positive, and R the mean radius of the Earth ($R = 6.371 \times 10^8$ cm).

By a cross differentiation, we have from (1) and (2)

$$-\frac{\partial}{\partial z} \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) + 2\omega \sin \varphi \left\{ \frac{\partial}{\partial x}(\rho u) + \frac{\partial}{\partial y}(\rho v) \right\} + \frac{2\omega \cos \varphi}{R} \rho v = 0. \quad (4)$$

Eliminating $\frac{\partial}{\partial x}(\rho u) + \frac{\partial}{\partial y}(\rho v)$ between (3) and (4), we have

$$-\frac{\partial}{\partial z} \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) + \frac{2\omega}{R \cos \varphi} \rho v - 2\omega \sin \varphi \frac{\partial}{\partial z}(\rho w) = 0. \quad (5)$$

Substitution from (1) in (5) gives

$$\frac{\partial p}{\partial x} = \frac{\partial}{\partial z} \left\{ \tau_x + R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) + 2\omega R \cos \varphi \sin^2 \varphi \rho w \right\}. \quad (6)$$

If we integrate both side of this equation from the surface down to a depth z , we have since ζ is small

$$\begin{aligned} \int_{-\zeta}^z \frac{\partial p}{\partial x} dz &= \int_0^z \frac{\partial p}{\partial x} dz = \left\{ \tau_x + R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \right. \\ &\quad \left. + 2\omega R \cos \varphi \sin^2 \varphi \left\{ \rho(z) w(z) - \rho_0 w_0 \right\} \right\} \end{aligned} \quad (7)$$

where ρ_0 , w_0 are the density and downward velocity at the surface, while, $\rho(z)$, $w(z)$ are those at a depth z . From this equation we have

$$\begin{aligned} -w(z) &= - \frac{\int_0^z \frac{\partial p}{\partial x} dz}{2\omega R \cos \varphi \sin^2 \varphi \rho(z)} \\ &\quad + \frac{\left\{ \tau_x + R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \right\} \Big|_0^z}{\rho \omega R \cos \varphi \sin^2 \varphi \rho(z)} - \frac{\rho_0}{\rho(z)} \cdot w_0. \end{aligned} \quad (8)$$

Now we have

$$dp = g \rho dz.$$

Integrating from the surface $z = -\zeta$ down to a depth $z = z$, we have

$$\begin{aligned} p - p_0 &+ \int_{-\zeta}^z g \rho dz \\ &= p_0 + g \rho \zeta + \int_0^z g \rho dz \end{aligned}$$

where p_0 is the atmospheric pressure at the sea surface $z=-\zeta$. Then we have

$$\frac{\partial p}{\partial x} = \frac{\partial p_0}{\partial x} + g\rho_0 \frac{\partial \zeta}{\partial x} + \int_0^z g \frac{\partial \rho(z)}{\partial x} dz. \quad (9)$$

Integrating again, we have

$$\begin{aligned} \int_{-\zeta}^h \frac{\partial p}{\partial x} dz &= \left(\frac{\partial p_0}{\partial x} + g\rho_0 \frac{\partial \zeta}{\partial x} \right) (h + \zeta) + \int_{-\zeta}^h dz \int_0^z g \frac{\partial \rho(z_1)}{\partial x} dz_1, \\ &= \left(\frac{\partial p_0}{\partial x} + g\rho_0 \frac{\partial \zeta}{\partial x} \right) h + \int_0^h g \frac{\partial \rho(z)}{\partial x} (h - z) dz \end{aligned} \quad (10)$$

neglecting small quantities. But we have from (6)

$$\begin{aligned} \int_{-\zeta}^h \frac{\partial p}{\partial x} dz &= \int_0^h \frac{\partial p}{\partial x} dz \\ &= \left\{ \tau_x + R \cos \varphi \sin^2 \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \right\} \Big|_0^h \\ &\quad + 2\omega R \cos \varphi \sin^2 \varphi (\rho_h w_h - \rho_0 w_0) \end{aligned} \quad (11)$$

where ρ_h and w_h are the density and vertical velocity of sea water at the bottom.

Equating the right-hand sides of (10) and (11), we have

$$\begin{aligned} \frac{\partial p_0}{\partial x} + g\rho_0 \frac{\partial \zeta}{\partial x} &= -\frac{1}{h} \int_0^h g \frac{\partial \rho(z)}{\partial x} (h - z) dz \\ &\quad + \frac{1}{h} \left\{ \tau_x + R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \right\} \Big|_0^h \\ &\quad + 2\omega R \cos \varphi \sin^2 \varphi \cdot \frac{1}{h} (\rho_h w_h - \rho_0 w_0) \end{aligned} \quad (12)$$

again neglecting small quantities.

Eliminating $\frac{\partial p_0}{\partial x} + g\rho_0 \frac{\partial \zeta}{\partial x}$ between (9) and (12), we have

$$\begin{aligned} \frac{\partial p}{\partial x} &= \int_0^z g \frac{\partial \rho(z)}{\partial x} dz - \frac{1}{h} \int_0^h g \frac{\partial \rho(z)}{\partial x} (h - z) dz \\ &\quad + \frac{1}{h} \left\{ \tau_x + R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \right\} \Big|_0^h \\ &\quad + 2\omega R \cos \varphi \sin^2 \varphi \cdot \frac{1}{h} (\rho_h w_h - \rho_0 w_0). \end{aligned} \quad (13)$$

Substituting this expression for $\frac{\partial p}{\partial x}$ in (8), we have

$$-w(z) = \frac{\int_0^z g \frac{\partial \rho(z_1)}{\partial z} (z - z_1) dz - \frac{z}{h} \cdot \int_0^h g \frac{\partial \rho(z_1)}{\partial x} (h - z_1) dz_1 + q(z)}{2\omega R \cos \varphi \sin^2 \varphi \rho(z)} \quad (14)$$

where

$$\begin{aligned} q(z) &= -\left\{ \tau_x - R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \right\} \Big|_0^z \\ &\quad + \frac{z}{h} \left\{ \tau_x - R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \right\} \Big|_0^h \\ &\quad + 2\omega R \cos \varphi \sin^2 \varphi \left\{ \left(1 - \frac{z}{h} \right) \rho_0 w_0 + \frac{z}{h} \rho_h w_h \right\} \end{aligned} \quad (15)$$

In the expression for $q(z)$ above we have

$$\tau_x - R \cos \varphi \sin \varphi \left(\frac{\partial \tau_y}{\partial x} - \frac{\partial \tau_x}{\partial y} \right) \leq 1$$

except for some particular areas, and

$$2\omega R \cos \varphi \sin^2 \varphi \zeta(z) < 3.6 \times 10^4$$

so that if we adopt for w_0 and w_h values much smaller than 10^{-3} , the usual values of upwelling at intermediate levels, the values of $q(z)$ can be estimated at

$$|q(z)| \leq 1$$

Except at levels very close to the surface and bottom, the upwelling velocity $-w(z)$ will be given by

$$-w(z) = \frac{\int_0^z g \frac{\partial \rho(z_1)}{\partial x} (z - z_1) dz_1 - \frac{z}{h} \cdot \int_0^h g \frac{\partial \rho(z_1)}{\partial x} (h - z_1) dz_1}{2\omega R \cos \varphi \sin^2 \varphi \cdot \rho(z)} \tag{16}$$

Table 1. Velocities of Upwelling (–) and Sinking (+) (cm/sec)
computed from POFI NorPac data.

depth (m)	Stations		
	81-83	83-84	84-86
0	0	0	0
10	-0.028×10^{-4}	-0.062×10^{-4}	$+0.005 \times 10^{-4}$
20	-0.054	-0.124	+0.008
30	-0.079	-0.186	+0.009
50	-0.121	-0.306	+0.009
75	-0.163	-0.443	+0.004
100	-0.199	-0.568	-0.009
150	-0.276	-0.822	-0.044
200	-0.309	-1.054	-0.118
250	-0.328	-1.281	-0.193
300	-0.346	-1.498	-0.257
400	-0.434	-1.942	-0.332
500	-0.432	-2.283	-0.450
600	-0.411	-2.554	-0.560
700	-0.392	-2.740	-0.645
800	-0.369	-2.845	-0.701
1000	-0.464	-3.050	-0.676
1200	-0.402	-2.771	-0.634
1500	-0.778	-3.492	-0.522
2000	-0.623	-2.793	-0.417
2500	-0.467	-2.095	-0.313
3000	-0.311	-1.397	-0.209
3500	-0.156	-0.698	-0.104
4000	0	0	0

This formula will not be valid at the equator and poles ($\varphi=0, \pm\frac{\pi}{2}$), because the denominator vanishes there. However, we may point out that the upwelling velocity increases toward the equator indefinitely but without changing signs when we cross the Equator.

The validity of this formula at the layers close to the surface and bottom depends upon how large the numerator of (16) is in comparison to $q(z)$ given by (15). We shall discuss this point by some numerical examples.

Table 1 gives a result of computation of vertical water movement by the formula (16). The negative and positive signs give the upwelling and sinking respectively. We used density data at four NorPac stations 81, 83, 84 and 86 of the Pacific Oceanic Fishery Investigations occupied on board the M/S Hugh M. SMITH on her cruise 30, July-August 1955 and published in the Special Scientific Report-Fisheries No. 168, U.S. Department of the Interior, Fish and Wildlife Service. These stations are located at approximately the same latitude 50°N . In the computation, we assumed

$$\frac{\partial \rho}{\partial x} \text{ and } \frac{\partial \rho}{\partial y}$$

vanish below 1000 m level. The result shows that there is weak upwelling in most depths.

Reference

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A Contribution to the Computation of Three-Dimensional Ocean Currents by High-Speed Computers

By

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Abstract

A method is suggested by which the distribution of three dimensional ocean currents in an assigned oceanic area may be computed with the use of high-speed computers. The velocity of currents will be computed from the observed distribution of pressure gradients in the sea and the stresses of winds at the sea surface. Effects of both the vertical and lateral eddy viscosity are taken into consideration, provided the values of their coefficients should be assumed.

1. Introduction

In recent development in dynamical oceanography treating the computation of oceanic circulation, there have been employed two different methods. On the first and older one use by STOMMEL (1947), MUNK (1950) and HIDAOKA (1951, 1955) the terms of pressure gradients are eliminated by a cross differentiation. The other method is to try to find out the velocity of current in terms of pressure gradients and wind stresses since the latter can be known from oceanographic observations.

Geostrophic approximations developed by Scandinavian scholars early this century can be involved in this latter category, since the expressions for the ocean currents are at once obtained in terms of pressure gradients by neglecting the terms of friction in the hydrodynamical equations. This method known as the "dynamical computation of ocean currents" has widely been used with big success, particularly for the levels excluding the thin surface and bottom layers more or less disturbed by friction.

Some of the solutions of this type were treated by V.W. EKMANN (1905), in the existence of friction, for a simple types of vertical distribution of pressure gradients and has been called the "convection currents of EKMANN's type", though comparatively little attention has been paid them by later oceanogra-

phers. EKMAN treated this type of currents independently from drift current and considered that the actual ocean currents consist of these several kinds of currents of different origins.

In two recent papers, HIDAOKA (1956, 1957) published a more general solution of this type in an integral form, solving the equations of stationary ocean currents for observed pressure gradients and wind stresses, subject to both surface and bottom conditions. This solution was obtained by retaining vertical eddy viscosity only in terms of friction.

In this paper, a method will be proposed to solve a more general problem taking both the vertical and horizontal eddy viscosity into account. The problem will give three-dimensional distribution of ocean currents. The solution in a closed form is not possible in this case, but we may be able preferably by the use of high-speed computers to obtain a numerical solution in terms of pressure gradients, wind stresses on the sea surface. Only difficulty is that we have to assign certain values to the coefficients of both vertical and horizontal mixing, as we have hitherto done in similar types of problems.

2. The Problem

Take the x -axis positive eastward, the y -axis northward and the z -axis vertically downward from the undisturbed level surface. Then the steady-state equations of motion are as follows:

$$\begin{cases} A_l \frac{\partial^2 u}{\partial x^2} + A_l \frac{\partial^2 u}{\partial y^2} + A_z \frac{\partial^2 u}{\partial z^2} + 2\omega\rho \sin \phi \cdot v = \frac{\partial p}{\partial x} \\ A_l \frac{\partial^2 v}{\partial x^2} + A_l \frac{\partial^2 v}{\partial y^2} + A_z \frac{\partial^2 v}{\partial z^2} - 2\omega\rho \sin \phi \cdot u = \frac{\partial p}{\partial y} \end{cases} \quad (1)$$

where u and v are the x -component and y -component of velocity respectively, p is the pressure, ρ the density of sea water, A_l the coefficient of horizontal mixing, A_z the coefficient of vertical mixing, ω the angular velocity of the Earth's rotation and ϕ the latitude. When we use the complex representation of velocity $\bar{w} = u + iv$, equations (1) become

$$A_l \frac{\partial^2 \bar{w}}{\partial x^2} + A_l \frac{\partial^2 \bar{w}}{\partial y^2} + A_z \frac{\partial^2 \bar{w}}{\partial z^2} - i \cdot 2\omega\rho \sin \phi \cdot \bar{w} = \frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y}. \quad (2)$$

For boundary conditions we take

$$\begin{cases} z=0; & -A_z \frac{\partial \bar{w}}{\partial z} = \tau \\ z=h; & \bar{w}=0 \end{cases} \quad (3)$$

where τ is the complex representation of wind-stress $\tau = \tau_x + i\tau_y$ and h is the depth to the bottom.

In addition to the surface and bottom conditions, we need a number of

boundary conditions at lateral boundaries.

The present problem now is to discuss how it is possible to compute three-dimensional ocean currents by integrating the hydrodynamical equations (1) as they stand, subject to the surface and bottom as well as some lateral boundary conditions.

3. Simplified Solutions

If we neglect all the frictional terms in (1) $A_l = A_z = 0$, we have the well known geostrophic currents:

$$u = -\frac{\partial p / \partial y}{2\omega\rho \sin \phi}; \quad v = \frac{\partial p / \partial x}{2\omega\rho \sin \phi}. \quad (4)$$

Further EKMAN's convection current can be obtained by solving the equations in which $A_l = 0$, $A_z = 0$ or

$$\left. \begin{aligned} A_z \frac{\partial^2 u}{\partial z^2} + 2\omega\rho \sin \phi v &= \frac{\partial p}{\partial x}; \\ A_z \frac{\partial^2 v}{\partial z^2} - 2\omega\rho \sin \phi u &= \frac{\partial p}{\partial y} \end{aligned} \right\} \quad (5)$$

or

$$A_z \frac{\partial^2 \bar{w}}{\partial z^2} - i \cdot 2\omega\rho \sin \phi \bar{w} = \frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \quad (6)$$

when $\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y}$ is given by a simple function of z , and subject to the conditions:

$$\left. \begin{aligned} z=0; \quad -A_z \frac{\partial \bar{w}}{\partial z} &= 0; \\ z=h; \quad \bar{w} &= 0 \end{aligned} \right\} \quad (7)$$

HIDAOKA's more general solution involves the terms of the observed pressure gradients and wind stresses. It is given by

$$\bar{w} = \frac{\tau_x + i\tau_y}{(1+i)aA_z} \cdot \frac{\sinh(1+i)a(h-z)}{\cosh(1+i)ah} - \frac{1}{A_z} \int_0^h \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_s G(z, s) ds \quad (8)$$

where $G(z, s)$ is a symmetric kernel of the type:

$$\left. \begin{aligned} G(z, s) &= \frac{\sinh(1+i)a(h-z+s) + \sinh(1+i)a(h-z-s)}{2(1+i)a \cosh(1+i)ah} \\ &\quad (s < z) \\ &= \frac{\sinh(1+i)a(h+z-s) + \sinh(1+i)a(h-z-s)}{2(1+i)a \cosh(1+i)ah} \\ &\quad (s > z) \end{aligned} \right\} \quad (9)$$

Afterwards, SAINT-GUILY (1959) succeeded in deriving the barometric and slope currents in addition to the drift and convection currents obtained by

HIDAKA, thus succeeding in deriving a complete set of ocean currents according to their origins.

In a paper by HIDAKA and NAGATA (1958), a further development was achieved by taking the lateral eddy viscosity into account. In this case, however, the problem was only confined to a flow of water parallel to EW-direction, so that it was reduced to a two-dimensional one and the variation of pressure gradients, wind stresses and velocity in EW-direction was neglected.

In the present paper, we are to solve a three-dimensional problem. There is no simplification of the problem given by the hydrodynamical equations (1), (2) and conditions (3).

4. Three-Dimensional Solution

To solve the equations (1), let us assume

$$u(x, y, z) = \sum_{s=1}^{\infty} u_s(x, y) \cos \frac{(2s-1)\pi z}{2h}; \quad (10)$$

$$v(x, y, z) = \sum_{s=1}^{\infty} v_s(x, y) \cos \frac{(2s-1)\pi z}{2h}; \quad (11)$$

$$\frac{\partial p}{\partial x} = \sum_{s=1}^{\infty} \left(\frac{\partial p}{\partial x} \right)_s \cos \frac{(2s-1)\pi z}{2h}; \quad \frac{\partial p}{\partial y} = \sum_{s=1}^{\infty} \left(\frac{\partial p}{\partial y} \right)_s \cos \frac{(2s-1)\pi z}{2h}; \quad (12)$$

there will be no difficulty in these expansions, provided we have

$$u_s(x, y) = \frac{2}{h} \int_0^h u(x, y, \zeta) \cos \frac{(2s-1)\pi \zeta}{2h} d\zeta; \quad (12)$$

$$v_s(x, y) = \frac{2}{h} \int_0^h v(x, y, \zeta) \cos \frac{(2s-1)\pi \zeta}{2h} d\zeta; \quad (14)$$

$$\left(\frac{\partial p}{\partial x} \right)_s = \frac{2}{h} \int_0^h \frac{\partial p}{\partial x} \cos \frac{(2s-1)\pi z}{2h} dz; \quad \left(\frac{\partial p}{\partial y} \right)_s = \frac{2}{h} \int_0^h \frac{\partial p}{\partial y} \cos \frac{(2s-1)\pi z}{2h} dz. \quad (15)$$

Further assume an expansion of the form:

$$A_z \frac{\partial^2 u}{\partial y^2} = \sum_{s=1}^{\infty} \cos \frac{(2s-1)\pi z}{2h} \cdot \frac{2}{h} \int_0^h A_z \frac{\partial^2 u}{\partial \zeta^2} \cos \frac{(2s-1)\pi \zeta}{2h} d\zeta;$$

then we have with Stokes

$$\frac{2}{h} \int_0^h A_z \frac{\partial^2 u}{\partial \zeta^2} \cos \frac{(2s-1)\pi \zeta}{2h} d\zeta = \frac{2}{h} \cdot \frac{\tau_x}{\rho} - \frac{(2s-1)^2 \pi^2 A_z}{4h^2 \rho} u_s(x, y)$$

so that

$$A_z \frac{\partial^2 u}{\partial z^2} = \sum_{s=1}^{\infty} \left\{ \frac{2}{h} \cdot \frac{\tau_x}{\rho} - \frac{(2s-1)^2 \pi^2 A_z}{4h^2 \rho} u_s(x, y) \right\} \cos \frac{(2s-1)\pi z}{2h}; \quad (16)$$

in a similar manner

$$A_z \frac{\partial^2 v}{\partial z^2} = \sum_{s=1}^{\infty} \left\{ \frac{2}{h} \cdot \frac{\tau_x}{\rho} - \frac{(2s-1)^2 \pi^2 A_z}{4h^2 \rho} v_s(x, y) \right\} \cos \frac{(2s-1)\pi z}{2h}. \quad (17)$$

These expansions are all possible when $z=h$ represents the bottom or some other level of no motion where both u and v vanish. Theoretically, these are infinite series, but in case of usual serial observations consisting of only n layers, the summation will extend only from $s=1$ to $s=n$.

Substitution of (10), (11), (12), (16) and (17) in (1) gives

$$\left. \begin{aligned} \frac{\partial^2 u_s}{\partial x^2} + \frac{\partial^2 u_s}{\partial y^2} - \frac{(2s-1)^2 \pi^2 A_z}{4h^2 A_l} v_s - \frac{2\omega \rho \sin \phi}{A_l} v_s &= \frac{1}{A_l} \left(\frac{\partial p}{\partial x} \right)_s - \frac{2}{h} \cdot \frac{\tau_x}{\rho A_z} \\ \frac{\partial^2 v_s}{\partial x^2} + \frac{\partial^2 v_s}{\partial y^2} - \frac{2(s-1)^2 \pi^2 A_z}{4h^2 A_l} u_s + \frac{2\omega \rho \sin \phi}{A_l} u_s &= \frac{1}{A_l} \left(\frac{\partial p}{\partial y} \right)_s - \frac{2}{h} \cdot \frac{\tau_x}{\rho A_z} \end{aligned} \right\} \quad (18)$$

($s=1, 2, \dots, n$)

Thus we could reduce the three-dimensional problem given by (1) and (3) to a two-dimensional problem, though the number of equations has increased.

5. Solution by Successive Approximations

Suppose we divide the domain under consideration by two systems of straight lines parallel to latitude circles and meridians. Let the distance between successive lines be d .

The two systems of equations (18) become on replacing the Laplacian operator $\frac{\partial^2}{\partial x^2} + \frac{\partial^2}{\partial y^2}$ by difference approximation, or

$$\left(\frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} \right)_{m,n} = \frac{u_{m+1,n} + u_{m-1,n} + u_{m,n+1} + u_{m,n-1} - 4u_{m,n}}{d^2} \quad (19)$$

where the left-hand member is the value of $\frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2}$ at a particular point (x_m, y_n) and $u_{m,n}$, $u_{m+1,n}$, $u_{m-1,n}$, $u_{m,n+1}$ and $u_{m,n-1}$ are the values of u at the points (x_m, y_n) , (x_{m+d}, y_n) , (x_{m-d}, y_n) , (x_m, y_{n+d}) and (x_m, y_{n-d}) respectively, where d is the distance between successive points, assumed constant in both meridional and EW-direction. This condition will be approximately fulfilled if the domain under consideration is coincident with a tropical belt whose boundaries are not very far from the equator.

$$\begin{array}{c} \bullet(m, n+1) \\ \bullet(m-1, n) \quad \bullet(m, n) \quad \bullet(m+1, n) \\ \bullet(m, n-1) \end{array}$$

The same also applies to v , or

$$\left(\frac{\partial^2 v}{\partial x^2} + \frac{\partial^2 v}{\partial y^2} \right)_{m,n} = \frac{v_{m+1,n} + v_{m-1,n} + v_{m,n+1} + v_{m,n-1} - 4v_{m,n}}{d^2} \quad (20)$$

Substituting (19) and (20) in (18), we have

$$\left\{4 + \frac{(2s-1)^2\pi^2}{4} \left(\frac{A_z}{A_l}\right) \left(\frac{d}{h}\right)^2\right\}_{s_1} u_{mn} = -\frac{1}{A_l} \left(\frac{\partial p}{\partial x}\right)_s - \frac{2}{h} \cdot \frac{\tau_x}{\rho A_l} + \frac{2\omega\rho \sin \phi_n}{A_l} v_{mn} + u_{m-1,n} + u_{m+1,n} + u_{m,n-1} + u_{m,n+1}, \quad (21)$$

$$\left\{4 + \frac{(2s-1)^2\pi^2}{4} \left(\frac{A_z}{A_l}\right) \left(\frac{d}{h}\right)^2\right\}_{s_1} v_{mn} = -\frac{1}{A_l} \left(\frac{\partial p}{\partial y}\right)_s - \frac{2}{h} \cdot \frac{\tau_y}{\rho A_l} - \frac{2\omega\rho \sin \phi_n}{A_l} u_{mn} + u_{n-1,n} + v_{m+1,n} + v_{m,n-1} + v_{m,n+1},$$

where ϕ_n is the latitude at the point (x_m, y_n) . The two above sets of equations are valid at all the grid points in the domain under consideration. It will be possible to solve these equations by successive approximations. From these equations, we can first assume that 0th approximations are $u_{m,n}^{(0)} = 0$, $v_{m,n}^{(0)} = 0$ for all points in the domain under consideration. By the use of successive approximation the $(\nu+1)$ th approximation on the left-hand side can be computed by substituting ν th approximations on the left-hand side members of the equations:

$$X_{mn}^{(\nu+1)} = \left\{ \begin{aligned} & -\frac{1}{A_l} \left(\frac{\partial p}{\partial x}\right)_s - \frac{2}{h} \frac{\tau_x}{\rho A_l} + \frac{2\omega\rho \sin \phi_n}{A_l} Y_{mn}^{(\nu)} + X_{m-1,n}^{(\nu)} + X_{m+1,n}^{(\nu)} + X_{m,n-1}^{(\nu)} + X_{m,n+1}^{(\nu)} \\ & \frac{4 + \frac{(2s-1)^2\pi^2}{4} \left(\frac{A_z}{A_l}\right) \left(\frac{d}{h}\right)^2}{\left\{4 + \frac{(2s-1)^2\pi^2}{4} \left(\frac{A_z}{A_l}\right) \left(\frac{d}{h}\right)^2\right\}_{s_1}} \end{aligned} \right\} \quad (22)$$

$$Y_{mn}^{(\nu+1)} = \left\{ \begin{aligned} & -\frac{1}{A_l} \left(\frac{\partial p}{\partial y}\right)_s - \frac{2}{h} \frac{\tau_y}{\rho A_l} - \frac{2\omega\rho \sin \phi_n}{A_l} X_{mn}^{(\nu)} + Y_{m-1,n}^{(\nu)} + Y_{m+1,n}^{(\nu)} + Y_{m,n-1}^{(\nu)} + Y_{m,n+1}^{(\nu)} \\ & \frac{4 + \frac{(2s-1)^2\pi^2}{4} \left(\frac{A_z}{A_l}\right) \left(\frac{d}{h}\right)^2}{\left\{4 + \frac{(2s-1)^2\pi^2}{4} \left(\frac{A_z}{A_l}\right) \left(\frac{d}{h}\right)^2\right\}_{s_1}} \end{aligned} \right\}$$

where X_{mn} and Y_{mn} represent u_{mn} and v_{mn} respectively.

The processes can be repeated until the values of $X_{m,n}$ and $Y_{m,n}$ become stationary all over the domain, the use of high-speed computer will be most recommended in carrying out the computations.

This iteration process always requests that a set of boundary values be given because the values of X_{mn} and Y_{mn} at the point (X_m, Y_m) can be improved always by employing their values at four surrounding points (x_m-d, y_n) , (x_m+d, y_n) , (x_m, y_m-d) and (x_m, y_m+d) , the boundary values being given in advance and remaining constant throughout the entire operation. The choice of these boundary values at first seems quite tiring because there will be no way to find them out. Indeed it is by no means possible to know them correctly.

Very fortunately, however, we shall be able to prove that we can put all these boundary values for u and v equal to zero. Of course, actual boundary

values are not zero, so that the result of computation will be more or less influenced by this simplification. Their influence is, however, only limited within narrow belts close to the boundaries involving a few grid points across these belts.

6. Proof that the Boundary Values of u and v can be = 0

It will be fortunate if we can put all the boundary values for u_s and v_s equal to zero, in computing them by iterations using the equations (22). The proof for this will be given in the following lines.

In equation (2) we replace the second derivative of $\bar{w}$ with respect to y with a finite difference approximation, or

$$\frac{\partial^2 \bar{w}}{\partial y^2} = \frac{\bar{w}_{n-1} - 2\bar{w}_n + \bar{w}_{n+1}}{d^2}$$

and suppose a quadrilateral ocean area whose boundaries are not the coasts. It may be bounded by water surface on all sides, or partly bounded by the lands. In order to simplify the problem let us assume the area bounded by two parallels $x=0, a$ and two meridians $y=0, b$. Let the values of $\bar{w}$ at the northern and southern boundaries of the domain under consideration be $\bar{w}_N$ and $\bar{w}_S$ respectively. Then equation (2) becomes the following sets of simultaneous equations:

$$\left. \begin{aligned} A_l \frac{d^2 \bar{w}_1}{dx^2} + A_z \frac{d^2 \bar{w}_1}{dz^2} + A_l \frac{\bar{w}_N - 2\bar{w}_1 + \bar{w}_2}{d^2} - i \cdot 2\omega\rho \sin \phi_1 \bar{w}_1 &= \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_1, \\ A_l \frac{d^2 \bar{w}_2}{dx^2} + A_z \frac{d^2 \bar{w}_2}{dz^2} + A_l \frac{\bar{w}_1 - 2\bar{w}_2 + \bar{w}_3}{d^2} - i \cdot 2\omega\rho \sin \phi_2 \bar{w}_2 &= \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_2, \\ &\dots\dots\dots \\ A_l \frac{d^2 \bar{w}_m}{dx^2} + A_z \frac{d^2 \bar{w}_m}{dz^2} + A_l \frac{\bar{w}_{m-1} - 2\bar{w}_m + \bar{w}_{m+1}}{d^2} - i \cdot 2\omega\rho \sin \phi_m \bar{w}_m &= \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_m \end{aligned} \right\} \quad (23)$$

Multiplying both sides of the equations (4) by undertermined multipliers $l_1, il_2, i^2 l_3, \dots, i^{k-1} l_k, \dots, i^{m-1} l_m$,

$$\left. \begin{aligned} Q(\sum_k i^{k-1} l_k \bar{w}_k) + \frac{A_l}{d^2} (l_1 \bar{w}_N + i^{m+1} l_m \bar{w}_S) \\ + \{ -(2 + i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_1) l_1 + il_2 \} \bar{w}_1 \\ + \{ l_1 - (2 + i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_2) il_2 + i^2 l_3 \} \bar{w}_2 \\ + \{ il_2 - (2 + i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_3) i^2 l_3 + i^3 l_4 \} \bar{w}_3 \end{aligned} \right\} \quad (24)$$

$$\left. \begin{aligned} &+ \dots\dots \\ &+ \{i^{m-2}l_{m-1} - (2+i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_m) i^{m-1}l_m\} \bar{w}_m \\ &= \sum_{k=1}^m i^{k-1} l_k \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_k \end{aligned} \right\}$$

where

$$Q(\bar{w}) = A_l \frac{\partial^2 \bar{w}}{\partial x^2} + A_z \frac{\partial^2 \bar{w}}{\partial z^2} \quad (25)$$

Now let

$$\left. \begin{aligned} &-(2+i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_1) l_1 + i l_2 - (2+i\xi) l_1, \\ &l_1 - (2+i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_2) i l_2 + i^2 l_3 = -(2+i\xi) i l_2, \\ &i l_2 - (2+i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_3) i^2 l_3 + i^3 l_4 = -(2+i\xi) i^2 l_3, \\ &\dots\dots\dots \\ &i^{m-2} l_{m-1} - (2+i \cdot 2\omega\rho \frac{d^2}{A_l} \sin \phi_m) i^{m-1} l_m = -(2+i\xi) i^{m-1} l_m, \end{aligned} \right\} \quad (26)$$

then we have

$$Q(V) - \frac{A_l}{d^2} (2+i\xi) V = -\frac{A_l}{d^2} (l_1 \bar{w}_N + i^{m-1} l_m \bar{w}_S) + \sum_{k=1}^m i^{k-1} l_k \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_k \quad (27)$$

where

$$\begin{aligned} V &= l_1 \bar{w}_1 + i l_2 \bar{w}_2 + i^2 l_3 \bar{w}_3 + \dots i^{m-1} l_m \bar{w}_m \\ &= \sum_{k=1}^m i^{k-1} l_k \bar{w}_k \end{aligned} \quad (28)$$

The equations (26) will be transformed into a form :

$$\left. \begin{aligned} &(\xi - 2\omega\rho \frac{d^2}{A_l} \sin \phi_1) l_1 + l_2 = 0, \\ &l_1 + (\xi - 2\omega\rho \frac{d^2}{A_l} \sin \phi_2) l_2 + l_3 = 0, \\ &l_2 + (\xi - 2\omega\rho \frac{d^2}{A_l} \sin \phi_3) l_3 + l_4 = 0, \\ &\dots\dots\dots \\ &l_{m-1} + (\xi - 2\omega\rho \frac{d^2}{A_l} \sin \phi_m) l_m = 0 \end{aligned} \right\} \quad (29)$$

If we solve these equations for ξ , we have m roots :

$$\xi_1, \xi_2, \xi_3, \dots, \xi_j, \dots, \xi_m$$

for each of which we have different sets of l_k or l_{kj} . Equation (28) then becomes

$$Q(V_j) - (2 + i\xi_j) - V_j = \frac{A_l}{d^2} (l_{1j}\bar{w}_N + i^{m-1}l_{mj}\bar{w}_S) + \sum_{k=1}^m i^{k-1}l_{kj} \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_k \quad (30)$$

where

$$V_j = \sum_{k=1}^m i^{k-1}l_{kj}\bar{w}_k; \quad \tau_j = \sum_{k=1}^m i^{k-1}l_{kj}\tau_k \quad (31)$$

The values of l_{kj} , or the values of $l_1, l_2, \dots, l_k, \dots, l_m$ for the j th of the m roots of (29) were determined numerically. In this case we selected a tropical belt extending from $11^\circ.5N$ to $11^\circ.5S$, the spacing of the successive stations being 1° of latitude. This makes $k=22$ excluding two stations on northern and southern boundaries. For $A_l=10^8$ c. g. s. which is considered to be a rather large value, we have the table I. In this table, we see that the l_{kj} are largest on the diagonal $k=j$ and become smaller as $|j-k|$ increases. For this reason we conclude that l_{1j} and l_{mj} are small when $j-1$ or $m-j$ are not large, or if we exclude one or two stations annexed to the boundaries, the terms.

$$\frac{A_l}{d^2} (l_{1j}\bar{w}_N + i^{m-1}l_{mj}\bar{w}_S)$$

are not very important for stations a few times d away from the boundaries, and can be neglected. This means that if we take the northsouth expanse of the equatorial belt large enough, we can put the boundary values of $\bar{w}$ or $\bar{w}_N$ and $\bar{w}_S$ equal to zero.

For smaller values of A_l the decrease of l_{kj} as we go away from the diagonal becomes much more pronounced. For this reasons, it is not necessary for us to take the values of $\bar{w}$ at the northern and southern boundaries into consideration.

By above simplification, equation (30) becomes

$$\begin{aligned} A_l \frac{\partial^2 V_j}{\partial x^2} + A_s \frac{\partial^2 V_j}{\partial z^2} - \frac{A_l}{d^2} (2 + i\xi_j) V_j &= \sum_{k=1}^m i^{k-1}l_{kj} \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_k \\ &= \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_j \\ (j=1, 2, \dots, m) \end{aligned} \quad (32)$$

Now assume with Stokes in a similar way as we did in § 4,

$$V_j(x, z) = \sum_{s=1}^{\infty} V_j^s(x) \cos \frac{(2s-1)\pi z}{2h} \quad (33)$$

and

$$\left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_j = \sum_{s=1}^{\infty} \left(\frac{\partial p}{\partial x} + i \frac{\partial p}{\partial y} \right)_j^s \cos \frac{(2s-1)\pi z}{2h} \quad (34)$$

where

$$V_j^s(x) = \frac{2}{h} \int_0^h V_j(x, \zeta) \cos \frac{(2s-1)\pi \zeta}{2h} d\zeta \quad (35)$$

Table 1. Values of the roots ξ_j of the equations (29) and corresponding values of l_{kj} for $A_t=10^8$ c. g. s.

j	ξ_j	Values of l_{kj}																						$10^{\circ}.5S$			
		$\varphi=10^{\circ}.5N$		$9^{\circ}.5$		$8^{\circ}.5$		$7^{\circ}.5$		$6^{\circ}.5$		$5^{\circ}.5$		$4^{\circ}.5$		$3^{\circ}.5$		$2^{\circ}.5$		$1^{\circ}.5$		$0^{\circ}.5N$			$1^{\circ}.5S$		
		$k=1$	2	3	4	5	6	7	8	9	10	11	12	13		21	22		21	22							
1	33.469	1	0.331	0.054	0.006	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
2	30.631	0.316	1	0.297	0.045	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
3	27.416	0.047	0.298	1	0.298	0.046	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
4	24.210	0.005	0.046	0.298	1	0.297	0.045	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
5	20.997	0	0.005	0.046	0.297	1	0.296	0.045	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
6	17.778	0	0	0.005	0.045	0.297	1	0.296	0.045	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
7	14.553	0	0	0	0.005	0.045	0.296	1	0.296	0.045	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0	0	etc.	etc.
8	11.323	0	0	0	0	0.005	0.045	0.296	1	0.295	0.045	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0
9	8.091	0	0	0	0	0	0	0	0.045	0.295	1	0.295	0.045	0.005	0	0	0	0	0	0	0	0	0	0	0	0	0
10	4.855	0	0	0	0	0	0	0	0	0.005	0.045	0.295	1	0.295	0.045	0.005	0	0	0	0	0	0	0	0.005	0	0	0
11	1.619	0	0	0	0	0	0	0	0	0	0.005	0.045	0.295	1	0.295	0.045	0.005	0	0	0	0	0	0.045	0.005	0	0	0
12	-1.619	0	0	0	0	0	0	0	0	0	0	0	0	0.005	0.045	0.295	1	0.295	0.045	0.005	0	0	0.045	0.005	0	0	0
13	-4.855	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0.005	0.045	0.295	1	0.295	0.045	0.005	0	0	0
.....																											
.....																											0.047
.....																											1 0.316
.....																											0.331 1
22	-33.469																										

and

$$\left(\frac{\partial p}{\partial x} + i\frac{\partial p}{\partial y}\right)_j^s = \frac{2}{h} \int_0^h \left(\frac{\partial p}{\partial x} + i\frac{\partial p}{\partial y}\right)_j \cos \frac{(2s-1)\pi z}{2h} dz. \quad (36)$$

These expansions will never be difficult if we consider both the current and pressure gradients vanish at a certain depth, or

$$z=h: V_j=0; \quad \left(\frac{\partial p}{\partial x} + i\frac{\partial p}{\partial y}\right)_j = 0. \quad (37)$$

The depth $z=h$ can be either regarded as the bottom of the sea itself, or an arbitrarily chosen motionless level in the computation of relative currents. At the sea surface, we shall have

$$\begin{aligned} z=0: -A_z \frac{\partial V_j}{\partial z} &= \sum_{k=1}^m i^{k-1} l_{kj} \tau_k \\ &= \tau_j \end{aligned} \quad (38)$$

Since the number of the observed levels is actually finite, we shall have

$$V_j(x, y) = \sum_{s=1}^n V_j^s(x) \cos \frac{(2s-1)\pi z}{2h}; \quad (39)$$

$$\left(\frac{\partial p}{\partial x} + i\frac{\partial p}{\partial y}\right)_j = \sum_{s=1}^n \left(\frac{\partial p}{\partial x} + i\frac{\partial p}{\partial y}\right)_j^s \cos \frac{(2s-1)\pi z}{2h}, \quad (40)$$

and the conditions to be satisfied by V_j are of course

$$z=0: -A_z \frac{\partial V_j}{\partial z} = \tau_j \quad (41)$$

and

$$z=h: V_j=0 \quad (42)$$

Now assume $A_z \frac{\partial^2 V_j}{\partial z^2}$ can be also expanded in a similar series, or

$$A_z \frac{\partial^2 V_j}{\partial z^2} = \sum_{s=1}^n \cos \frac{(2s-1)\pi z}{2h} \cdot \frac{2}{h} \int_0^h A_z \frac{\partial^2 V_j}{\partial \zeta^2} \cos \frac{(2s-1)\pi \zeta}{2h} d\zeta;$$

then we have by Stokes' method

$$A_z \frac{\partial^2 V_j^s}{\partial z^2} = \sum_{s=1}^n \left\{ \frac{2}{h} \cdot \frac{\tau_j}{\rho} - \frac{(2s-1)^2 \pi^2 A_z}{4h^2 \rho} V_j^s(x) \right\} \cos \frac{(2s-1)\pi z}{2h}. \quad (43)$$

Substitution of (33), (34) and (43) into (32) gives

$$A_i \frac{d^2 V_j^s}{dx^2} - \left\{ \frac{A_i}{d^2} (2 + i\xi_j) + \frac{(2s-1)^2 \pi^2 A_z}{4h^2 \rho} \right\} V_j^s = - \frac{2}{h\rho} \tau_j + \left(\frac{\partial p}{\partial x} + i\frac{\partial p}{\partial y} \right)_j^s. \quad (44)$$

If we put

$$b_j^s = + \sqrt{\frac{(2s-1)^2 \pi^2 A_z}{4h^2 \rho A_i} + (2 + i\xi_j) \frac{1}{d^2}} \quad (45)$$

$$F_j^s(x) = -\frac{2}{h\rho}\tau_j + \left(\frac{\partial p}{\partial x} + i\frac{\partial p}{\partial y}\right)_j^s \quad (46)$$

equation (44) will become

$$\frac{d^2 V_j^s}{dx^2} - (b_j^s)^2 V_j^s = -\frac{1}{A_l} F_j^s(x). \quad (47)$$

If we denote the boundary values of V_j^s at the western and eastern boundaries or $x=0$, a , as

$$\left. \begin{aligned} x=0: V_j^s &= W_j^s \\ x=a: V_j^s &= E_j^s \end{aligned} \right\} \quad (48)$$

Solution of (47) subject to the conditions (48) is

$$V_j^s(x) = \frac{\sinh b_j^s x}{\sinh b_j^s a} E_j^s + \frac{\sinh b_j^s (a-x)}{\sinh b_j^s a} W_j^s - \frac{1}{A_l} \int_0^a F_j^s(\sigma) K_j^s(x, \sigma) d\sigma \quad (49)$$

where $K_j^s(x, \sigma)$ is a symmetric kernel of the type:

$$\begin{aligned} K_j^s(x, \sigma) &= \frac{\sinh b_j^s (a-x) \sinh b_j^s \sigma}{b_j^s \sinh b_j^s a} \quad (\sigma < x) \\ &= \frac{\sinh b_j^s (a-\sigma) \sinh b_j^s x}{b_j^s \sinh b_j^s a} \quad (\sigma > x) \end{aligned} \quad (50)$$

Now the solution (49) contains the boundary value W_j^s and E_j^s with factors

$$\frac{\sinh b_j^s x}{\sinh b_j^s a} \quad \text{and} \quad \frac{\sinh b_j^s (x-a)}{\sinh b_j^s a} \quad (51)$$

and the third term on the right-hand member of (49) involves no boundary values. The first function in (51) is only significant for a value of x very close to $x=a$, or at the eastern boundary. The second quantity in (51) on the contrary is only significant at a domain very close to the western boundary or $x=0$. Thus except the narrow belts very close to the western and eastern boundary, the terms in the solution (49) involving boundary values can be neglected. Thus we shall have

$$V_j^s(x) = -\frac{1}{A_l} \int_0^a F_j^s(\sigma) K_j^s(x, \sigma) d\sigma \quad (52)$$

independent of the boundary values provided x is close to neither 0 nor a . Thus the solution (52) applies to almost all parts of the domain under consideration.

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Equatorial Flow and Inertia Terms

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Abstract

In this paper, is present a method to compute the velocity of currents exactly at the equator, taking into account the field accelerations in addition to the pressure gradients and Coriolis forces alone. The result was numerically compared with HIDAHA-TSUCHIYA's formulae 1955, which were derived from the consideration of a simple geostrophic balance only.

1. Introduction

Several years ago, TSUCHIYA (TSUCHIYA, 1955) and HIDAHA (HIDAHA, 1955) almost simultaneously proposed a method for the computation of currents in a dynamical section across the equator where the previous method of geostrophic balance has failed at the equator due to the existence of $\sin$ (latitude) in the denominator of the expression.

In this paper, this problem was checked again, taking the non-linear terms due to the field accelerations in the dynamical equations into account. Some numerical results were worked out and compared with those computed from the formulae of HIDAHA-TSUCHIYA 1955.

2. Non-linear Equations for Equatorial Flows.

Suppose a meridional section passing through the equator. Take the x -, y - and z -axis positive eastward, northward and vertically downward, and let the components of velocity in these directions be u , v w and respectively. Assume that the east-west variation of velocity components are small compared with the meridional, and neglect the terms $\frac{\partial u}{\partial x}$ and $\frac{\partial v}{\partial x}$. Further disregarding the thermodynamic effects, the frictional terms due to eddy viscosity, and the terms $w \frac{\partial u}{\partial z}$ and $w \frac{\partial v}{\partial z}$ in field accelerations, the steady state equations of ocean currents become

$$\left. \begin{aligned} 2\omega \sin \varphi v - v \frac{\partial u}{\partial y} &= \frac{1}{\rho} \frac{\partial p}{\partial x}; \\ -2\omega \sin \varphi u - v \frac{\partial v}{\partial y} &= \frac{1}{\rho} \frac{\partial p}{\partial y}. \end{aligned} \right\} \quad (1)$$

In these expressions, p is the pressure, ρ the density of sea water, ω the angular velocity of the Earth and φ the geographic latitude connected with y by a relation $R \cdot \varphi = y$ where R is the radius of the Earth. Further the equations (1) can be transformed as

$$\left. \begin{aligned} 2\omega \sin \varphi v - v \frac{\partial u}{\partial y} &= \frac{\partial D}{\partial x}; \\ -2\omega \sin \varphi u - v \frac{\partial v}{\partial y} &= \frac{\partial D}{\partial y}. \end{aligned} \right\} \quad (2)$$

where D is the geopotential distance of an isobaric surface referred to a motionless layer which is chosen sufficiently deep.

Very close to the equator, the slope of an isobaric surface will be given by

$$\left. \begin{aligned} \frac{\partial D}{\partial x} &= d_0 + d_1 y + d_2 y^2; \\ \frac{\partial D}{\partial y} &= \delta_0 + \delta_1 y + \delta_2 y^2 \end{aligned} \right\} \quad (3)$$

respectively, where d_0, d_1, d_2 and $\delta_0, \delta_1, \delta_2$ are constants, and can be determined from the results of hydrographic observations.

3. Solutions

To solve the equations (2) in which the quantities $\frac{\partial D}{\partial x}$ and $\frac{\partial D}{\partial y}$ are given by (3), we suppose that u , and v are respectively given by the expressions:

$$\left. \begin{aligned} u &= u_0 + u_1 y + \frac{u_2}{2} y^2; \\ v &= v_0 + v_1 y + \frac{v_2}{2} y^2, \end{aligned} \right\} \quad (4)$$

where $u_0, u_1, u_2; v_0, v_1, v_2$ are constants to be determined. Close to the equator, we shall have very accurately,

$$2\omega \sin \varphi = \frac{2\omega}{R} y. \quad (5)$$

Substitutions of (4) in (2) and putting to zero of several terms gives

$$u_1 v_0 + d_0 = 0, \quad (6)$$

$$\frac{2\omega}{R}v_0 - u_1v_1 - u_2v_0 - d_1 = 0, \quad (7)$$

$$\frac{2\omega}{R}v_1 - u_2v_1 - \frac{1}{2}u_1v_2 - d_2 = 0, \quad (8)$$

$$v_1v_0 + \delta_0 = 0, \quad (9)$$

$$-\frac{2\omega}{R}u_0 - v_1^2 - v_0v_2\delta_1 = 0, \quad (10)$$

$$-\frac{2\omega}{R}u_1 - \frac{3}{2}v_1v_2 - \delta_2 = 0 \quad (11)$$

Eliminating u_2 and v_2 between the equations (7), (8) and (10), we have

$$3u_1v_1^3 + \frac{2\omega}{R}u_1^2v_0 + 3d_1v_1^2 - 3d_2v_0v_1\delta_2u_1v_0 = 0. \quad (12)$$

On the other hand, we have from (6) and (9)

$$u_1 = -\frac{d_0}{v_0}; \quad v_1 = -\frac{\delta_0}{v_0}.$$

Substituting these in (12), we have

$$(3d_2\delta_0 - d_0\delta_2)v_0^4 + \frac{2\omega}{R}d_0^2 \cdot v_0^3 + 3d_1\delta_0^2 \cdot v_0^2 + 3d_0\delta_0^3 = 0 \quad (13)$$

This quartic equation will give as one of its real roots the value of v_0 in terms of d_0 , d_1 , d_2 , δ_0 , and δ_2 . Then it is not difficult to compute the expressions for u_0 , u_2 and v_2 in terms of d_0 , d_1 , d_2 ; δ_0 , δ_1 , δ_2 and v_0 . Thus we have

$$\left. \begin{aligned} u_0 &= -\frac{2d_2}{d_0} \cdot \frac{R}{2\omega} \cdot v_0^3 - \frac{2d_1\delta_0}{d_0} \cdot \frac{R}{2\omega} \cdot v_0 - \delta_1 \cdot \frac{R}{2\omega} - 3\frac{\delta_0^2}{v_0^2} \cdot \frac{R}{2\omega} \\ u_1 &= -\frac{d_0}{v_0}, \\ u_2 &= \frac{2\omega}{R} - \frac{d_1}{v_0} - \frac{d_0\delta_0}{v_0^3}, \\ v_1 &= -\frac{\delta_0}{v_0}, \\ v_2 &= \frac{2d_2}{d_0}v_0 + \frac{2d_1\delta_0}{d_0v_0} + \frac{2\delta_0^2}{v_0^3}. \end{aligned} \right\} \quad (14).$$

These 5 constants together with v_0 which is given as one of real roots of (13) will determined the meridional variations of the zonal and meridional components of velocity.

At the equator $\varphi=0$ or $y=0$, we have

$$\left. \begin{aligned} u &= -\frac{2d_2}{d_0} \cdot \frac{R}{2\omega} \cdot v_0^3 - \frac{2d_1\delta_0}{d_0} \cdot \frac{R}{2\omega} v_0 - \delta_1 \frac{R}{2\omega} - 3\frac{\delta_0^2}{v_0^2} \cdot \frac{R}{2\omega}, \\ v &= v_0 \end{aligned} \right\} \quad (15)$$

These expressions enable us to compute the velocity of flows exactly at the equator in terms of the slopes of isobaric surface.

4. Constant Slopes.

When the sloped of the isobaric surface are constant at the equator, or

$$\frac{\partial D}{\partial x} = d_0; \quad \frac{\partial D}{\partial y} = \delta_0 \quad (16)$$

we have $d_1 = d_2 = \delta_1 = \delta_2 = 0$ and the equation (13) becomes a cubic equation:

$$\frac{2\omega}{R} d_0^2 \cdot v_0^3 + 3d_0 \cdot \delta_0^3 = 0$$

from which we have

$$\left. \begin{aligned} v_0 &= -\delta_0 \left(-\frac{3}{d_0} \cdot \frac{R}{2\omega} \right)^{\frac{1}{3}}; \\ u_0 &= - \left(3d_0^2 \cdot \frac{R}{2\omega} \right)^{\frac{1}{3}}. \end{aligned} \right\} \quad (17)$$

and from (15)

These are expressions for an equatorial flow, when the slope of an isobaric surface is uniform around the equator.

5. Comparison with Hidaka-Tsuchiya's Expressions 1955.

HIDAKA-TSUCHIYA's expressions for current velocity at the equator was derived from the expressions

$$\left. \begin{aligned} 2\omega \sin \varphi v - \frac{\partial D}{\partial x} &= 0; \\ -2\omega \sin \varphi u - \frac{\partial D}{\partial y} &= 0, \end{aligned} \right\} \quad (18)$$

which are obtained by neglecting the inertia terms in (1) or (2). Since $\sin \varphi$ becomes identifiinitely small at the equator $\varphi=0$, these two authors advise to differentiate both members of (18) with respect to $y(=R \cdot \varphi)$ and then make $\varphi=0$ or $y=0$. Then we have the velocity components at the equator:

$$\left. \begin{aligned} u &= -\frac{(\partial^2 D / \partial y^2) \varphi = 0}{2\omega / R}, \\ v &= -\frac{(\partial^2 D / \partial y \partial x) \varphi = 0}{2\omega / R} \end{aligned} \right\} \quad (19)$$

provided either or both of $\frac{\partial^2 D}{\partial y^2}$ and $\frac{\partial^2 D}{\partial x \partial y}$ exist at $\varphi=0$. Substituting from (3), (19) become

$$u_0 = -\frac{R}{2\omega} \cdot \delta_1; \quad v_0 = -\frac{R}{2\omega} \cdot d_1 \quad (20)$$

and the result depends upon neither d_0 or δ_0 . The result has no concern with the values of the quantities d_0 and δ_0 . On the other hand, we have, if we have take field accelerations into account,

$$\left. \begin{aligned} u_0 &= -\frac{2d_1\delta_0}{d_1} \cdot v_0 \cdot \frac{R}{2\omega} - \delta_1 \cdot \frac{R}{2\omega} - 3\frac{\delta_0^2}{v_0^2} \cdot \frac{R}{2\omega} \\ v_0 &= -\delta_0 \cdot \left(\frac{3}{d_0} \cdot \frac{R}{2\omega} \right)^{\frac{1}{3}} \end{aligned} \right\} \quad (21)$$

providing $d_2=\delta_2=0$. The latter quantities need not necessarily vanish, but the expressions will be simpler when they do.

As is evident from a comparison of the expressions (20) and (21) HIDAHA-TSUCHIYA's formula 1955 and the present formulae derived under the consideration of the field accelerations are of entirely different natures.

6. Numerical Results.

In order to exemplify the comparison, let us apply both expressions to the hydrographic data at the stations occupied by TOWNSEND CROMWELL (CROMWELL 1951, 1954), THOMAS AUSTIN (AUSTIN, 1954) and E. D. STROUP (STROUP, 1954). Their reports give the data occupied at stations along several meridional sections across the equator, each station therein being located at 0° (equator), 1° N and S, 2° N and S, etc, or at a constant spacing 1° of meridional arc.

In order to remove small fluctuations appearing in the meridional variation of $\frac{\partial D}{\partial y}$, the anomalies of D over 1000 decibar surface in several meridional sections observed between 155° W and the International Date lines were averaged for the same latitudes. $\frac{\partial D}{\partial y}$ were then computed by making successive differences of D in meridional direction, and then dividing them by twice the length 111.12×10^5 cm or the length of 1° of meridional arc. Further, the value of $\frac{\partial D}{\partial y}$ at the middle points or 0.5° N and S, 1.5° N and S, etc were then computed by interpolation. Since however, the resulting curves themselves have still some small fluctuations around the equator, they were again smoothed out by the the rule:

$$\frac{1}{3} \left\{ \left(\frac{\partial D}{\partial y} \right)_{n-1} + \left(\frac{\partial D}{\partial y} \right)_n + \left(\frac{\partial D}{\partial y} \right)_{n+1} \right\}$$

where specifies a particular station under consideration of $\frac{\partial D}{\partial y}$ is shown in following Table I.

The values of $\frac{\partial D}{\partial x}$ were computed by HIDAHA and NAGATA (HIDAHA and NAGATA, 1958) only for the equator also from the Hawaiian observational results.

Table I. Slopes of Isobaric Surface $\frac{\partial D}{\partial y}$ and $\frac{\partial D}{\partial x}$ at and close to the Equator.

Pressure in decibars (approx. depth in m)	$\frac{\partial D}{\partial y}$ (in 10^{-5})							$\frac{\partial D}{\partial x}$ (in 10^{-5})
	1.5°S	1°S	0°.5S	0°	0°.5N	1°N	1°.5N	0°
0	-37.8	-15.3	+7.2	-4.1	-15.3	+8.6	+32.4	-3.96
20	-35.1	-14.0	+7.2	-4.1	-15.3	+5.9	+27.0	-3.63
40	-34.2	-12.2	+9.9	-4.5	-18.9	+1.8	+22.5	-3.16
60	-31.5	-11.3	+9.0	-4.5	-18.0	-1.4	+15.3	-2.82
80	-29.7	-10.4	+9.0	-5.0	-18.9	-2.7	+13.5	-2.51
100	-27.9	-8.6	+10.8	-5.0	-20.7	-6.3	+8.1	-2.33
120	-22.5	-6.8	+9.0	-5.4	-19.8	-9.0	+1.8	-1.79
140	-17.1	-3.6	+9.9	-5.9	-21.6	-11.7	-1.8	-1.24

The meridional variation of $\frac{\partial D}{\partial x}$ near the equator was then determined assuming

$$\frac{\partial D}{\partial y} = \delta_0 + \delta_1 y,$$

in order to meet the condition $\delta_2=0$ under which the expressions (21) were derived. The quantities δ_0 and δ_1 were then determined by the method of least squares. Since the values of $\frac{\partial D}{\partial x}$ were available at the equator only we have inevitably to put $d_1=d_2=0$. As a matter of fact, the east-west variation of $\frac{\partial D}{\partial x}$ is considered to be quite small, and we need not anticipate any disregarding d_1 and d_2 . The result of determination of these quantities is given in the following Table II.

Table II. Determination of δ_0 , δ_1 and d_0

depth (m)	δ_0	δ_1	d_0
0	-4.05×10^{-5}	-20.25×10^{-12}	-3.96×10^{-5}
20	-4.05	-20.25	-3.63
40	-4.50	-25.92	-3.16
60	-4.50	-24.30	-2.82
80	-4.95	-25.11	-2.51
100	-4.95	-28.35	-2.33
120	-5.40	-25.92	-1.79
140	-5.85	-28.35	-1.24

Substituting the values of δ_0 , δ_1 and d_0 in the expressions (21) with $d_1=0$, and in (20), we have the following table which clearly shows very remarkable

difference between the east-west component by two different theories.

Table III. Comparison of Equatorial Velocity Components Computed by the Present Theory with these Computed by Hidaka-Tsuchiya's Formula 1955.

depth (m)	Present Theory 1961		Hidaka-Tsuchiya's 1955	
	u_0	v_0	u_0	v_0
0	+61.07	-28.01	+88.46	—
20	+62.62	-28.84	+88.46	—
40	+89.67	-33.56	+113.23	—
60	+84.31	-34.85	+106.16	—
80	+89.49	-39.86	+109.70	—
100	+104.62	-40.86	+123.85	—
120	+97.10	-48.67	+113.23	—
140	+111.22	-59.59	+123.85	—

Discussion for the results of comparison will be left for later study.

7. Acknowledgements

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Sea Surface Temperature, Radiation, and Wind Anomalies in the North Pacific Ocean*

By

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Abstract

Nonseasonal variations of sea surface temperature occur in many parts of the ocean. These anomalies cover large areas and have an average endurance of three to six months. Equally large scale, but less persistent, variations occur in the sea level atmospheric pressure which may be interpreted as geostrophic wind anomalies, and in radiation. An examination of the relation of temperature anomalies to those of wind and radiation is made by computing the crosscorrelations. Significant values were found in some cases, but are at best restricted to certain times of the year and to certain areas.

Introduction

Sea surface temperature anomalies in almost any part of the world are of some interest to oceanographers, meteorologists and fisheries biologists. Various authors have sought to relate to them such matters as the frequency of hurricane occurrence (RIEHL, 1954), agricultural problems (HANZAWA and INOUE, 1952) and the anomalous distribution of organisms (HUBBS, 1948). Yet the extent and causes of sea surface temperature anomalies are only incompletely understood. Two possible causes are wind and radiation. The effect of wind upon temperature has received some attention, including studies by HELLAND-HANSEN and NANSEN (1920) and RODEWALD (1953) in the Atlantic, and HANZAWA (1957), HATANAKA (1952), HIRANO (1957) and UDA (1955) in the western Pacific. The influence of radiation has not been studied in detail owing to the difficulty of obtaining reliable data from oceanic regions.

In this paper we shall consider the dependence of the temperature anomalies upon anomalies of radiation and also of wind acting through advection and mixing. The importance of mixing in influencing the surface temperature depends critically upon the vertical temperature structure; if the temperature

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does not decrease appreciably with depth, i. e., the mixed layer is quite deep, as is frequently the case in winter in higher latitudes, mixing is not likely to have an appreciable influence upon the surface temperature. The anomalies are then principally the result of horizontal advection. If, on the other hand, the thermocline is shallow, mixing becomes increasingly important in altering the surface temperature.

Since there are not sufficient data available to study the depth of the mixed layer and the intensity of the underlying gradient, it is not possible to go further into the vertical transport of heat. Instead, much of the following discussion is limited to the winter season in temperate and high latitudes where there is reason to assume that the surface temperature change due to vertical heat transport is small compared to that due to horizontal movement. The strongest temperature anomalies can then be expected in regions where the wind blows across strong horizontal temperature gradients such as off northern Japan and in the region of the North Pacific westerlies.

Data

Monthly mean sea surface temperature records from 1921-1938 for the North Pacific Ocean were obtained from the publications of the Imperial Marine Observatory, Kobe, Japan. Sea surface temperature for 1957 and 1958 for the North Pacific east of longitude 180 were obtained from the Bureau of Commercial Fisheries, Honolulu Biological Laboratory in Hawaii. Monthly mean sea level atmospheric pressure charts for the period 1899-1958 (prepared from daily maps) were obtained from the U. S. Weather Bureau. Where necessary, pressure differences were converted into geostrophic winds. Geostrophic winds were not determined for ocean areas south of latitude 20 N, owing to the scarcity of pressure observations and the doubtful application of the geostrophic assumption in those places. From the geostrophic wind components the vertical component of the curl of the wind stress was determined.

Incoming short wave radiation was computed from cloudiness data (Imperial Marine Observatory, Kobe, Japan, 1921-1938) making use of the relation $Q = (1-r) Q_0(1-kn)$ where Q is the incoming radiation with a cloudy sky, r the amount reflected from the sea surface, Q_0 the incoming radiation with a clear sky, k a constant and n cloudiness in tenths. Values for r , Q_0 and k are given by BUDYKO (1956).

In all cases differences were taken between the monthly mean and the long term mean for the same month in order to eliminate the seasonal variation.

Relation between pressure and temperature anomalies

A qualitative agreement between pressure (wind) and temperature anomalies is seen by comparing the pressure anomaly maps with the correspond-

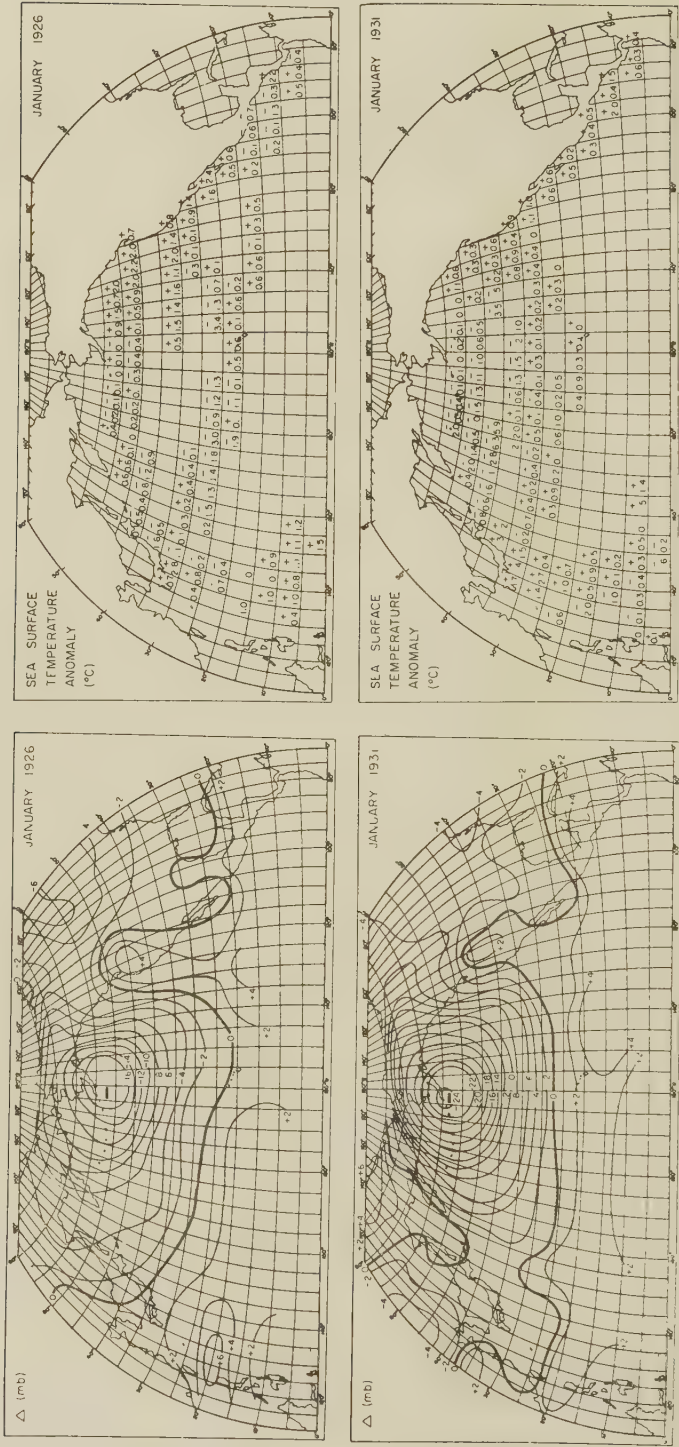


Fig. 1. Pressure and temperature anomalies during January 1926 and January 1931.

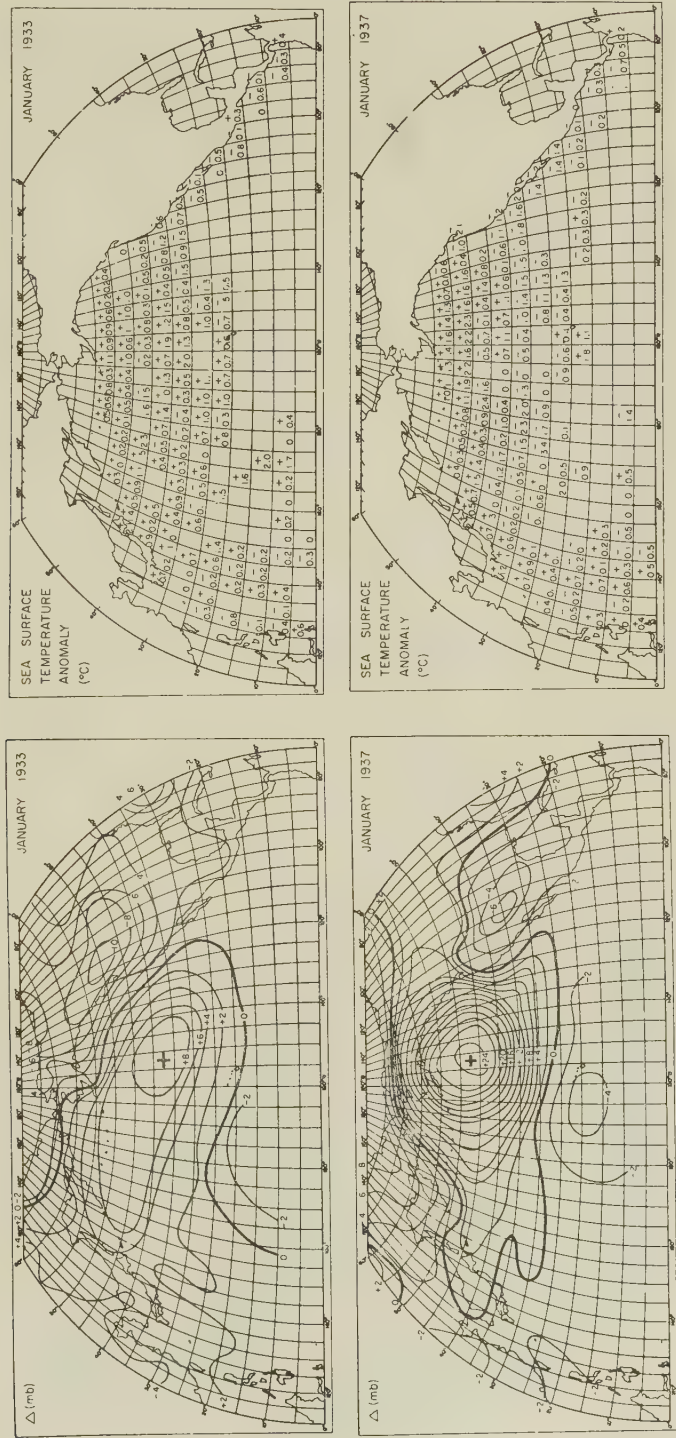


Fig. 2. Pressure and temperature anomalies during January 1933 and January 1937.

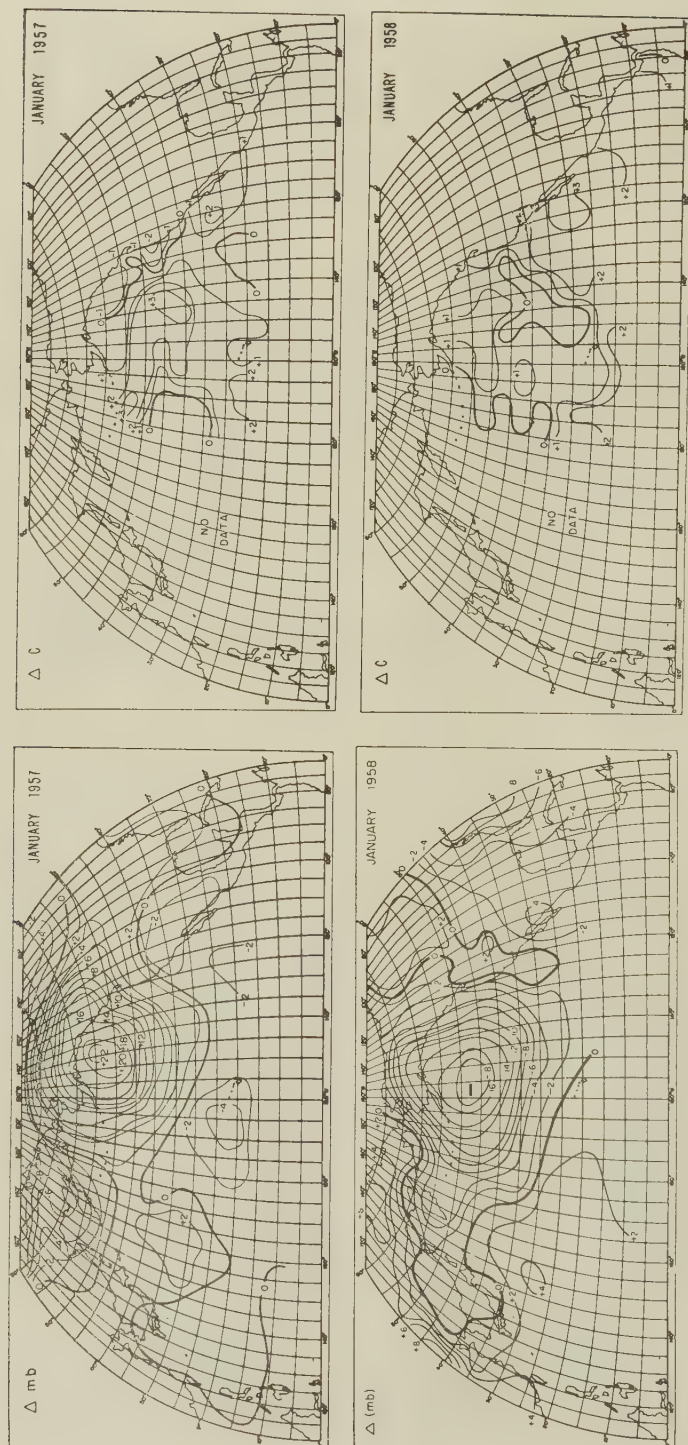


Fig. 3. Pressure and temperature anomalies during January 1957 and January 1958

ing temperature anomaly maps in Figs. 1, 2 and 3. Thus in January 1926, 1931 and 1958 the dominant feature in the central North Pacific was a large negative pressure anomaly. Since this anomaly occurs in the same region as the Aleutian Low, the circulation around this pressure cell was strengthened and consequently one could expect increased northerly winds in the northwestern and increased southerly winds in the Northeastern Pacific. If it is assumed that the surface waters move approximately in the direction of the wind or slightly to the right of it, one might anticipate negative temperature anomalies to the left and positive temperature anomalies to the right of the pressure anomaly. The observed temperature anomalies in January 1926, 1931 and 1958 show below average temperatures in the western central Pacific and above average temperatures in the eastern Pacific, as was expected from the above considerations. During January 1933, 1937 and 1957 the central North Pacific was occupied by a pronounced positive pressure anomaly. This anomaly was caused in 1933 by a strongly developed subtropical high pressure cell and in 1937 and 1957 by a southward extension of a polar anticyclone into regions normally occupied by the Aleutian Low. In the first case one could expect a reduced circulation around the Aleutian Low and in the last cases the atmospheric circulation was actually reversed. Thus, owing to the reduced northerly winds in the central and western Pacific, positive temperature anomalies could be expected there. In the eastern Pacific one would anticipate negative temperature anomalies owing to the reduced advection of warm water from the southwest. The observed temperature anomalies seem to confirm this.

The above examples shown were chosen to represent the strongest anomalies, but their large scale nature is a characteristic feature and not dependent upon the particular years selected. Notable negative pressure anomalies in the central Pacific occurred also in 1912 and 1941 and strong positive anomalies in 1911 and 1949. No temperature anomalies for the open ocean are available for these years but coastal records of air temperature (U.S. Weather Bureau) and sea surface temperature (U.S. Coast and Geodetic Survey, 1956) seem to indicate that 1912 and 1941 were warm and 1911 and 1949 cold winters along the shores of Alaska and the United States.

The increased or decreased circulation around the Aleutian Low can also be expressed in terms of anomalies in the curl of the geostrophic wind stress (Fig. 4), henceforth referred to as curl anomalies. Normally cyclonic (positive) curl prevails in the region of the Aleutian Low. Its magnitude is in the order of $3 \cdot 10^{-8}$ dynes cm^{-3} . (The mean January curl chart is remarkably similar to the chart of divergence and convergence areas in the Pacific Ocean in northern winter, prepared by HIDAOKA and OGAWA (1958)). In years when the circulation around the low is greatly increased the curl may be as high as $10 \cdot 10^{-8}$ dynes cm^{-3} and at times when a high pressure occupies the area of

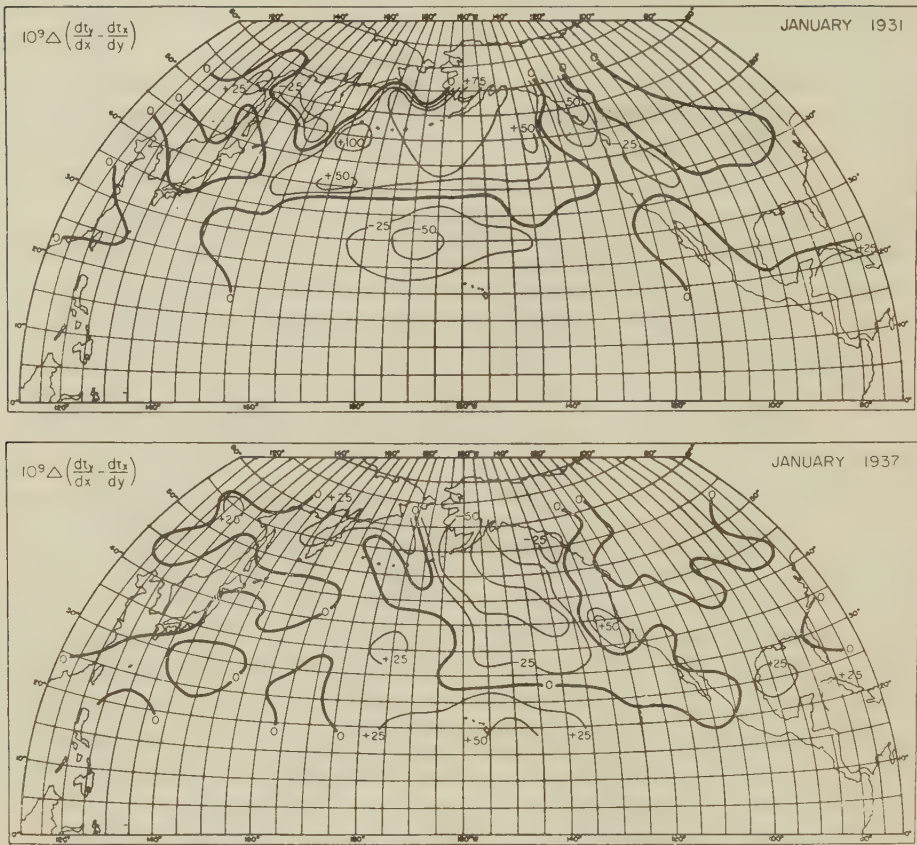


Fig. 4. Geostrophic wind stress vorticity anomalies for January 1931 and January 1937.

the Aleutian Low anticyclonic (negative) curl of the order of $2 \cdot 10^{-8}$ dynes cm^{-3} is found. The notable feature is the large scale nature of the anomalies and the fact that the anomaly is frequently larger than the mean. There is a very rough agreement between the respective areas of curl anomalies and temperature anomalies. If the respective charts for 1931 and 1937 are compared to each other, it is seen that cyclonic curl anomalies are accompanied by negative temperature anomalies in approximately the same region and that areas of anticyclonic curl anomalies correspond roughly to areas of positive temperature anomalies. Since both the cyclonic curl (through upwelling) and horizontal advection anomalies could have produced the negative temperature anomalies in the central Pacific, it is not clear which factor is the more important one. However, the small vertical temperature gradients normally observed in winter in high latitudes are suggestive that horizontal advection probably has the greater effect upon changing the surface temperatures there.

**Areal coherence, persistence and variance of temperature,
wind and radiation anomalies**

The areal coherence of temperature anomalies is illustrated for the California Current region in Fig. 5. It is seen that anomalies of the same sign and approximately of the same magnitude occur almost simultaneously between latitudes 20°N and 40°N.

The persistence of these anomalies can be calculated from the autocorrelation function :

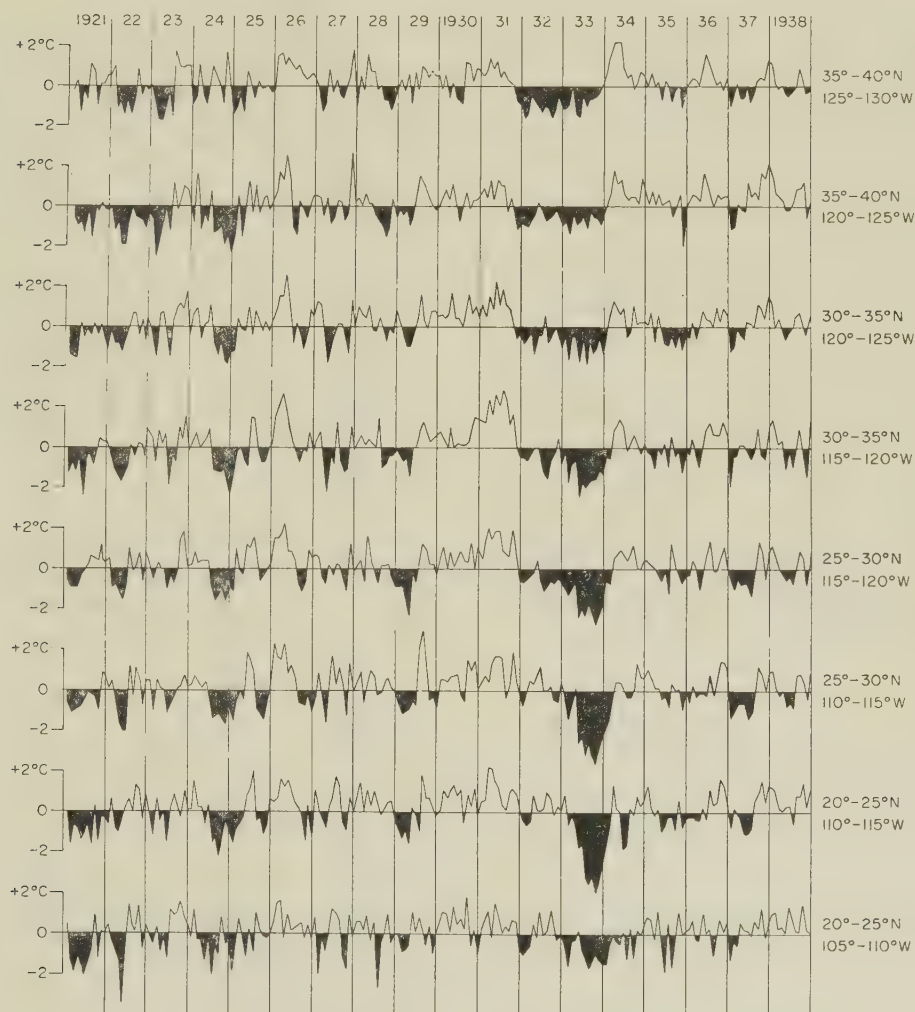


Fig. 5. Time series of temperature anomalies 1921-1938 for eight adjacent five degree squares in the California Current region.

$$\phi_{XX}(\tau) = \frac{\langle X(t) X(t+\tau) \rangle}{\langle X^2 \rangle}$$

where $\langle \rangle$ denotes the average taken over the record length and τ denotes the time lag in months. The standard deviation (σ) of the autocorrelation function (ϕ_{XX}) over the record length T can be estimated from the variance (σ^2) inequality (LANING and BATTIN, 1956).

$$\sigma^2 \leq \frac{4}{T} \sum_{\tau=1}^T \phi_{XX}^2(\tau)$$

The autocorrelation of temperature in the California Current region and several other regions in the North Pacific is shown in Fig. 6. The dashed line indicates the standard deviation defined above. It is seen that in widely different places the decay rates of the autocorrelations are similar and that temperature anomalies appear to be persistent, on an average, from three to six months. A notable exception is found east of Hokkaido in the Oyashio region where the temperature anomalies are slightly more persistent than elsewhere.

This particular period of persistence (three to six months) suggests at first that the anomalies might be expressed as minor variations of amplitude or phase from the mean annual curve. A shift in phase or amplitude of a sine curve of 12-month period would certainly result in an anomaly lasting six months followed by another six-month anomaly of opposite sign, that is, if we assume the anomaly during a given year could be written

$$X(t) = (a_0 + a \sin \omega t) - (b_0 + b \sin (\omega t + \varphi))$$

where a_0 is the mean of all months of all years, a the mean amplitude of seasonal variation, b_0 and b the corresponding values for a given year, and φ the phase shift from the mean during that year. The autocorrelation of the above expression is

$$\phi_{XX}(\tau) = (a_0 - b_0)^2 + \frac{1}{2}(a^2 + b^2 - 2ab \cos \varphi) \cos \omega \tau$$

The first term is a constant and, if considered alone, would give a persistence of a year. Anomalies of a year's duration are occasionally seen, but do not appear to be the dominating type of anomaly. From the second term one can see that an amplitude shift alone or phase shift alone lead to cosine curves with maxima and minima six months apart. If the autocorrelations are normalized and compared with those of Fig. 6, two striking differences are noted. First the cosine autocorrelation (which is a cosine curve of unity and period 12 months) does not decay so rapidly in the first two months as the observed autocorrelations; then the cosine autocorrelations have a minimum at six months and a maximum at twelve months, which the observed autocorrelations do not show. From these differences one may conclude that no systematic variations in phase and amplitude (large enough to be discerni-

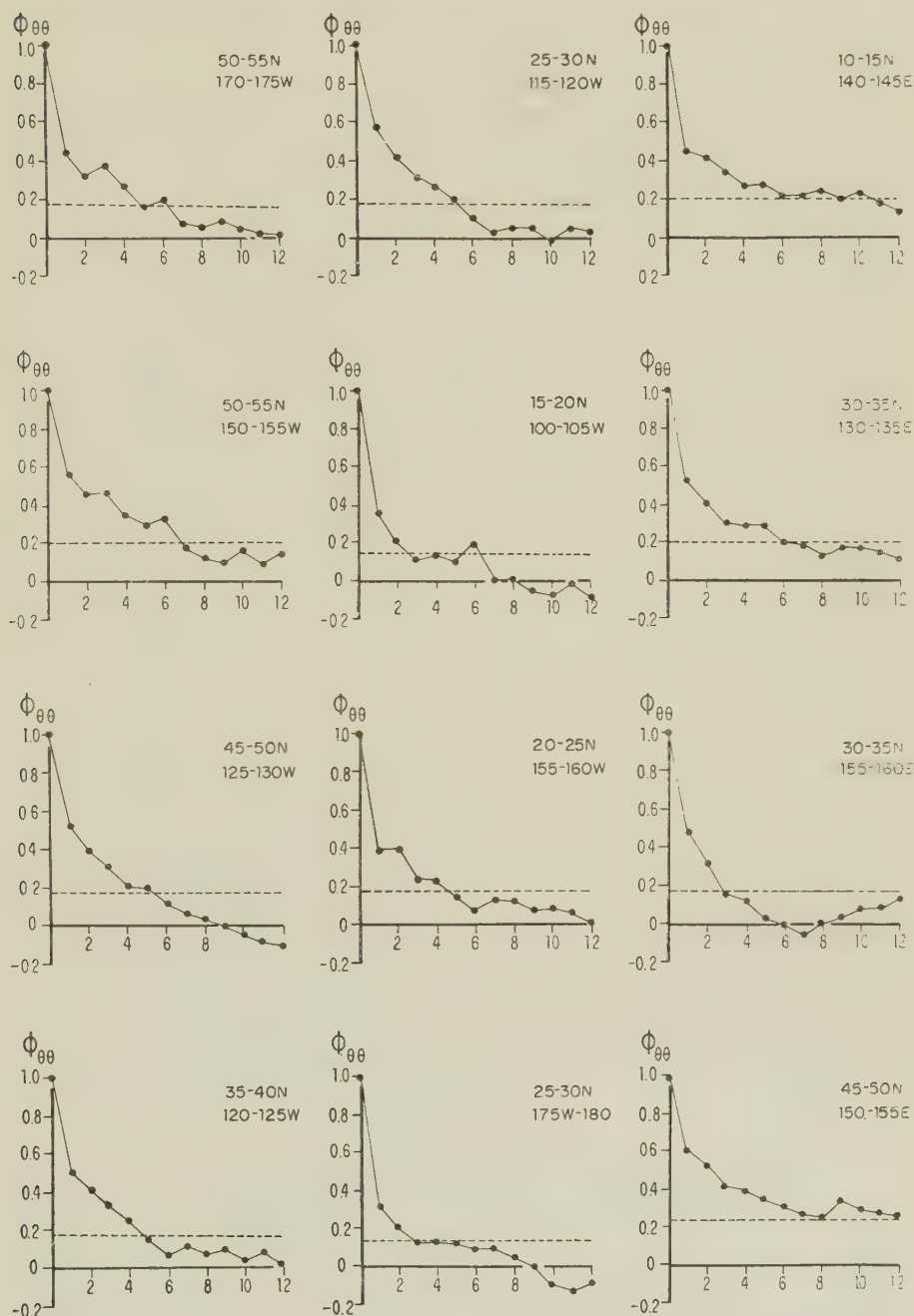


Fig. 6. Autocorrelations of temperature anomalies over the period 1921-1938 for lags from zero to twelve months. The dashed line indicates the standard deviation.

ble in the data over more than six months) have contributed substantially to the autocorrelations.

It is of interest to examine how the yearly mean, amplitude and phase have varied at a particular place. In

Table 1. Amplitude (A), phase (ϕ) and yearly mean (m) of sea surface temperatures ($^{\circ}\text{C}$) in area C 25-30N, 115-120W. σ denotes standard deviation.

Year	A $^{\circ}\text{C}$	ϕ deg.	m $^{\circ}\text{C}$
1921	2.8	214	17.5
1922	2.8	208	17.3
1923	2.3	202	17.6
1924	1.2	236	17.2
1925	2.1	233	17.8
1926	0.9	232	18.2
1927	2.0	218	17.5
1928	1.7	216	17.5
1929	3.0	221	17.3
1930	2.2	213	18.2
1931	2.0	232	18.6
1932	2.0	220	17.1
1933	1.7	213	15.9
1934	2.4	221	17.8
1935	2.2	215	17.4
1936	2.3	223	17.7
1937	2.9	209	17.2
1938	1.2	221	17.6
mean	2.1	219	17.5
σ	± 0.6	± 9	± 0.6

Table 1 these variations are shown for a five degree square off the coast of Baja California. It is seen that the standard deviation of phase is only nine degrees (days). The standard deviations of amplitude and annual mean are 0.6°C . Thus it appears very unlikely that the observed shifts in the seasons have any important effect upon the autocorrelations.

Similar investigations of the autocorrelations of radiation and wind anomalies led to the result that they were rarely persistent for more than a month (RODEN and GROVES, 1960).

The above autocorrelations were determined over the entire record length. If, on the other hand, the month to month variations of the autocorrelations are investigated, a different picture results. Nine ocean areas were chosen (Fig. 7) to study the month to month variations of temperature, wind and radiation anomalies.

The ocean areas are designated by capital letters and represent five degree squares of latitude and longitude for which temperature and radiation records were used. The east-west wind component (u) and the north-south wind component (v) were selected adjacent to each area and were computed from the pressures at the end points of the dotted lines shown in Fig. 7. For areas D and F winds were not available.

The monthly variation of the persistence of the anomalies is indicated by the variation of the autocorrelation between two adjacent months in Table 2. Significant values are in heavy print. The general impression gained from this table is that only for temperature anomalies is there a significant correlation from one month to the next. The few meaningful values in the autocorrelations of radiation and wind anomalies do not follow any simple pattern.

The variance of the anomalies is given in Table 3. It is seen that the variance of temperature anomalies varies very little with month. Over most

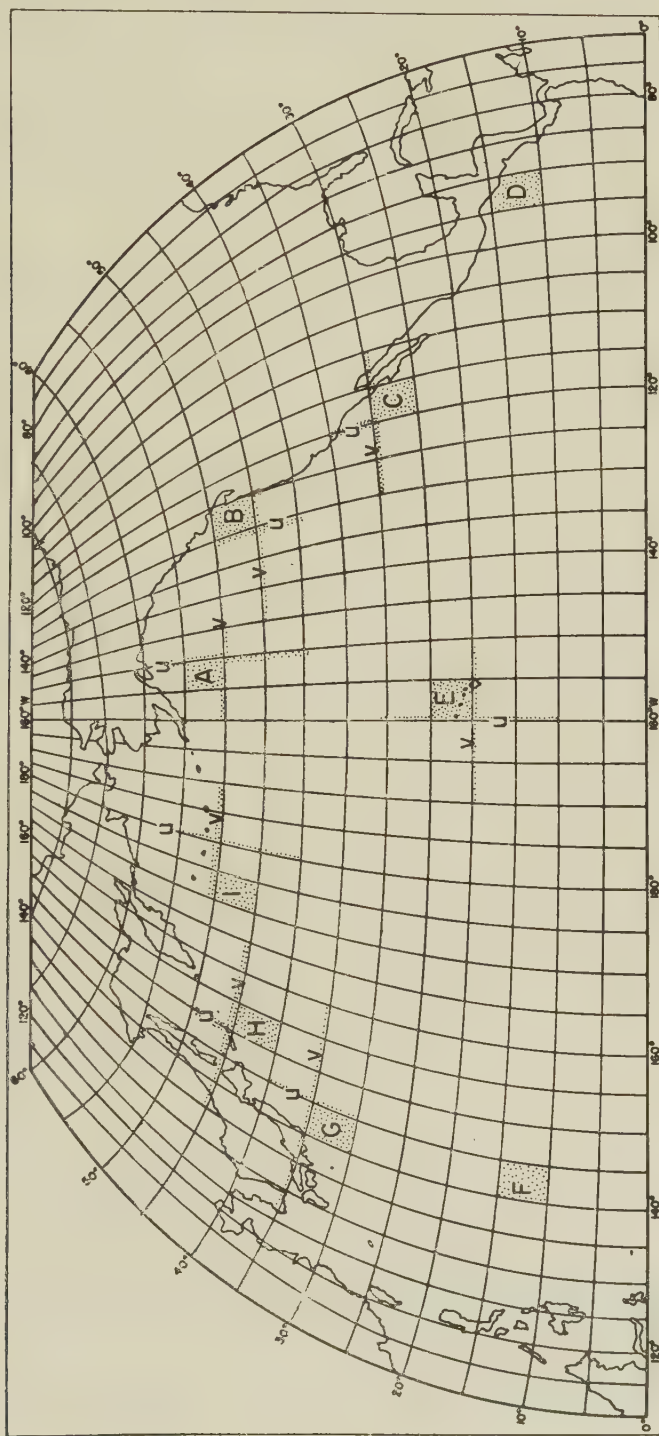


Fig. 7. Location of five-degree squares (A to I) in which temperature and radiation and corresponding east-west (u) and north-south (v) wind components are compared.

Table 2. Autocorrelation of temperature $\phi_{\theta\theta}$, radiation ϕ_{qq} and wind component ϕ_{uu} , ϕ_{vv} (u east, v north) anomalies between the present and next following month.

95% confidence limits are at ± 0.47 . Significant values are in heavy print.

	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	mean annual
$\phi_{\theta\theta}$													
A	0.54	-0.12	0.34	0.31	0.95	0.86	0.72	0.52	0.64	0.66	0.48	0.66	0.55
B	0.59	0.83	0.54	0.45	0.63	0.79	0.36	0.44	0.07	0.44	0.68	-0.03	0.48
C	0.78	0.60	0.89	0.52	0.80	0.67	0.34	0.80	0.74	0.24	0.25	0.60	0.60
D	0.47	0.44	0.05	-0.57	-0.36	0.31	0.34	0.82	0.38	0.52	0.54	0.41	0.27
E	0.30	0.09	0.55	0.42	0.20	0.67	0.36	0.54	0.30	0.40	0.49	0.47	0.40
F	0.59	0.23	0.47	0.34	0.89	0.46	0.58	0.27	0.54	0.46	0.41	0.05	0.44
G	0.27	0.94	0.28	0.81	0.75	0.42	0.31	-0.04	0.02	0.37	0.58	0.42	0.42
H	0.53	0.69	0.53	0.86	0.51	0.60	0.70	0.74	0.53	0.58	0.50	0.56	0.61
I	0.69	0.30	0.80	0.44	0.73	0.53	0.69	0.59	0.69	0.34	0.37	0.60	0.56
ϕ_{qq}													
A	0.13	0.65	0.53	-0.47	-0.19	0.17	-0.25	0.34	0.26	0.20	-0.12	0.37	0.14
B	-0.17	-0.55	0.33	0.08	0.08	-0.03	0.06	-0.55	-0.10	0.06	-0.02	-0.02	-0.07
C	-0.23	-0.03	0.88	-0.61	0.17	0.20	0.03	0.85	0.04	0.01	0.20	-0.26	0.10
D	0.16	-0.21	0.36	-0.56	-0.38	0.00	0.22	-0.02	-0.47	-0.11	0.02	0.08	-0.07
E	0.78	0.03	0.44	0.21	0.67	0.71	-0.18	0.08	-0.05	-0.22	0.11	-0.67	0.16
F	0.22	0.18	-0.14	-0.34	0.16	0.34	-0.32	-0.21	-0.11	-0.39	-0.08	0.26	-0.04
G	0.16	-0.16	0.76	0.09	0.52	0.05	0.14	-0.12	0.10	0.12	-0.17	0.49	0.17
H	0.28	-0.21	0.19	-0.28	-0.28	0.12	0.30	0.18	0.12	-0.06	0.00	0.38	0.06
I	-0.01	-0.25	0.47	0.16	-0.01	-0.25	-0.20	-0.34	0.08	0.02	-0.02	-0.09	-0.05
ϕ_{uu}													
A	-0.16	-0.13	0.05	0.05	0.69	-0.14	0.63	-0.17	0.60	0.21	0.10	0.00	0.14
B	-0.02	-0.27	0.33	0.02	0.04	-0.15	0.12	0.00	0.22	0.50	-0.14	0.26	0.08
C	-0.07	-0.17	-0.41	-0.15	0.11	0.30	0.00	-0.84	0.18	0.14	0.00	-0.31	-0.10
E	0.14	0.03	0.40	0.20	0.16	0.05	0.20	0.61	0.06	0.16	0.46	0.45	0.24
G	0.12	-0.15	0.26	-0.16	-0.15	-0.22	0.19	0.54	-0.04	-0.43	0.18	-0.18	-0.03
H	-0.23	-0.06	0.07	-0.03	-0.43	-0.02	-0.20	0.11	0.09	-0.06	-0.01	-0.15	-0.08
I	0.25	-0.37	0.01	0.03	0.14	-0.18	0.24	0.87	0.44	0.70	0.02	0.02	0.20
ϕ_{vv}													
A	0.03	-0.02	0.06	-0.35	0.07	0.58	-0.05	-0.22	0.27	0.18	-0.29	-0.30	0.00
B	0.33	0.09	0.36	0.32	-0.32	0.20	0.26	-0.30	-0.08	0.09	0.11	0.82	0.16
C	-0.13	0.36	0.30	0.41	0.30	0.25	-0.40	-0.50	0.47	-0.43	-0.80	-0.32	-0.04
E	0.22	0.19	0.45	0.22	0.23	0.27	0.20	0.17	0.46	0.06	0.05	0.23	0.23
G	-0.80	-0.02	0.05	-0.20	0.56	0.50	-0.06	-0.14	0.14	0.27	0.30	0.23	0.20
H	0.09	-0.24	-0.26	0.00	0.18	-0.31	-0.18	-0.27	0.25	0.31	-0.05	-0.12	-0.05
I	0.05	-0.01	-0.20	-0.58	0.11	0.33	0.11	-0.28	-0.49	0.10	0.28	0.36	-0.18

Table 3. Variance of temperature θ , radiation q and wind component (u east, v north) anomalies.

	I	II	III	VI	V	VI	VII	VIII	IX	X	XI	XII	mean annual
	<θ^2> ($^{\circ}\text{C}$)²												
A	0.8	0.7	0.3	0.5	0.2	0.6	0.8	1.0	0.7	1.1	0.8	0.7	0.7
B	0.5	0.6	0.8	1.1	0.6	0.6	0.6	0.4	0.4	0.4	0.4	0.6	0.6
C	0.6	0.8	0.8	1.0	0.9	0.8	0.1	0.6	0.8	1.0	0.9	0.5	0.8
D	0.5	0.6	0.7	0.7	0.4	0.5	0.4	0.3	0.3	0.3	0.2	0.3	0.4
E	0.9	0.4	0.8	0.3	0.4	0.7	0.8	0.3	0.2	0.4	0.6	0.6	0.5
F	0.5	0.6	0.7	0.7	0.4	0.5	0.4	0.3	0.3	0.3	0.2	0.3	0.4
G	0.9	0.4	0.8	0.3	0.4	0.7	0.8	0.3	0.2	0.4	0.6	0.6	0.5
H	2.0	2.2	2.1	1.3	1.4	1.0	1.5	2.4	2.6	1.3	2.0	2.1	1.8
I	0.3	0.4	0.4	0.3	0.3	0.5	0.6	0.7	0.7	0.7	0.5	0.4	0.5
	<q^2> (k cal cm⁻² mo⁻¹)²												
A	0.1	0.1	0.3	0.9	2.4	0.8	0.5	0.8	0.5	0.3	0.1	0.1	0.6
B	0.1	0.2	0.7	1.8	1.6	1.1	1.9	1.1	1.2	0.2	0.2	0.1	0.9
C	1.1	2.4	2.4	3.8	5.6	4.0	2.0	1.1	5.7	3.4	1.0	0.6	2.8
D	0.4	0.6	1.0	1.2	2.7	2.5	1.7	1.3	1.7	2.2	1.0	0.4	1.4
E	1.2	3.2	4.2	3.9	1.2	1.0	2.6	3.6	2.2	0.8	1.7	0.7	2.1
F	0.9	0.6	0.8	0.9	1.1	1.0	1.7	1.9	2.0	0.6	1.0	0.4	1.1
G	1.0	0.6	1.4	3.1	1.9	2.3	3.4	2.3	2.1	1.3	0.8	0.4	1.7
H	0.1	0.1	0.3	0.8	0.9	0.8	0.3	0.8	0.4	0.2	0.2	0.1	0.4
I	0.1	0.2	0.5	0.6	0.4	0.3	0.4	0.3	0.5	0.2	0.1	0.1	0.3
	<u^2> (m/s)²												
A	6.0	4.4	5.0	2.1	1.1	2.1	0.4	0.9	2.1	3.8	2.8	10.3	3.4
B	7.2	6.3	4.8	3.4	2.7	1.1	1.9	0.7	2.8	2.2	4.3	4.3	3.5
C	2.4	6.4	2.1	2.2	0.5	0.6	0.9	0.3	1.2	2.3	2.7	1.8	2.0
E	8.2	7.2	5.0	3.1	1.4	1.1	0.6	1.0	2.8	3.1	2.6	8.2	3.7
G	3.2	2.6	3.0	5.0	2.3	3.1	5.6	3.8	3.1	1.6	4.0	5.3	3.6
H	4.1	5.1	3.3	3.8	2.1	2.9	2.8	1.9	3.7	3.6	3.7	3.4	3.4
I	5.6	5.8	4.9	5.6	1.9	1.2	2.1	1.2	2.3	1.4	3.1	10.2	3.8
	<v^2> (m/s)²												
A	9.5	10.5	9.7	4.6	4.1	2.5	3.2	3.5	4.6	8.5	5.4	6.4	6.0
B	17.6	8.4	5.8	2.6	2.3	3.0	1.1	1.8	2.0	2.4	4.1	7.9	4.9
C	1.3	2.6	1.1	2.2	1.7	1.1	0.3	0.3	0.8	0.8	0.5	1.7	1.2
E	1.0	1.2	0.5	1.3	0.7	0.6	0.6	0.3	0.6	0.9	0.1	1.9	0.9
G	0.8	2.3	1.1	0.6	0.5	0.4	0.9	0.8	0.8	0.6	0.6	0.7	0.8
H	3.6	2.1	1.5	1.3	0.9	0.7	1.2	0.6	0.7	0.9	2.2	3.2	1.6
I	9.7	4.7	6.1	2.9	3.2	3.2	2.7	3.9	3.4	5.0	2.8	5.2	4.4

of the ocean it is less than 1 C², but a notable exception occurs off northern Japan in the boundary region between the Kuroshio and Oyashio where the variance is about twice as large.

The variance of radiation anomalies shows a characteristic maximum in summer and a minimum in winter. In areas with a semipermanent cloud-

cover, such as in the region of the North Pacific westerlies, the variance is smaller than in areas where the sky is only partly obscured by clouds, such as off the coast of Baja California. The mean value of the radiation variance is about $1.5 \text{ (kcal cm}^{-2} \text{ mo}^{-1})^2$.

The variance of wind anomalies shows a very pronounced seasonal variation with maximum values occurring in winter and minimum values in summer. The seasonal variation is particularly conspicuous in high latitudes and appears to be larger for the north-south components than for the east-west components. In middle latitudes, however, the east-west component is more variable. On an annual average the variance of the east-west component is not far from $3 \text{ m}^2 \text{ sec}^{-2}$, but for the north-south component it varies between $1 \text{ m}^2 \text{ sec}^{-2}$ and $6 \text{ m}^2 \text{ sec}^{-2}$.

Crosscorrelations between temperature, wind and radiation anomalies

The crosscorrelations between temperature and radiation and temperature and the two wind component anomalies are given in Table 4. The crosscorrelations were computed from the expression

$$\phi_{XY}(\tau) = \frac{\langle X(t) Y(t+\tau) \rangle}{\sqrt{\langle X^2 \rangle \langle Y^2 \rangle}}$$

for zero lag. Consideration of lag times other than zero was not possible owing to the short record length available.

It is seen that a few significant correlations between temperature and radiation anomalies do exist and that they are found mainly in the summer and early fall months. The summer and fall correlations are positive meaning that increased radiation leads to higher temperatures. The negative winter correlations in areas *D* and *F*, if real, are difficult to explain.

The crosscorrelations between wind and temperature show significant values mostly in the colder half of the year. It is noteworthy that an increased east to west component of the wind in the region of the North Pacific westerlies decreases the temperature in all significant cases. In the north-eastern Pacific, off the coast of Washington, increased southerly winds in winter are related to higher temperatures (probably through an increase in the Davidson Current). Off the coast of Baja California increased northerly winds are related to lower temperatures owing to increased upwelling and advection of cold water from the north (REID, RODEN, WYLLIE, 1958). In the Japanese region significant correlations between wind and temperature occur only in winter where stronger north winds lead to lower temperatures, as can be expected. In the Hawaiian region there appears to be no meaningful relation between the wind and temperature anomalies, possibly because of the doubtful application of the geostrophic relationship between pressure gradient and wind.

Table 4. Crosscorrelations between temperature and radiation $\phi_{\theta q}$ anomalies, temperature and east wind $\phi_{\theta u}$ anomalies and temperature and north wind anomalies $\phi_{\theta v}$. 95% confidence limit is at ± 0.47 .

Significant values are in heavy print.

	I	II	III	IV	V	VI	VII	VIII	IX	X	XI	XII	mean annual
$\phi_{\theta q}$													
A	-0.08	-0.06	0.46	-0.07	0.16	0.25	0.48	-0.10	-0.04	0.27	0.31	0.24	0.15
B	0.34	0.25	0.11	0.09	0.34	0.24	0.07	0.28	-0.08	0.05	0.19	-0.27	0.13
C	0.36	0.22	0.12	0.18	0.22	0.42	0.33	0.49	0.60	0.48	0.12	0.19	0.30
D	-0.32	-0.21	-0.46	-0.27	0.40	0.34	0.56	0.63	0.62	0.27	-0.22	-0.40	0.08
E	0.41	0.81	-0.03	-0.32	0.13	0.32	0.14	0.38	-0.44	0.10	0.33	0.15	0.16
F	0.24	-0.18	-0.59	0.45	-0.45	-0.01	0.28	-0.11	-0.05	0.20	0.11	-0.06	0.02
G	0.01	-0.34	0.25	0.08	-0.03	-0.33	-0.25	0.10	-0.04	-0.48	0.15	0.20	-0.06
H	0.04	0.04	-0.05	-0.13	0.02	0.15	0.56	0.28	0.00	0.22	0.29	-0.31	-0.09
I	0.19	-0.19	-0.10	-0.20	0.13	0.02	-0.01	0.58	-0.12	-0.17	-0.28	0.07	-0.01
$\phi_{\theta u}$													
A	-0.04	-0.46	-0.61	-0.25	-0.34	-0.66	-0.05	-0.19	-0.38	-0.40	0.03	0.30	-0.25
B	-0.09	0.08	-0.55	-0.48	-0.29	-0.05	0.35	-0.08	0.04	-0.19	0.08	0.25	-0.08
C	0.21	0.33	-0.53	0.22	0.11	-0.09	-0.24	-0.13	0.06	-0.31	0.10	-0.07	-0.03
E	0.13	-0.20	-0.08	0.34	0.28	-0.38	-0.34	0.33	0.18	-0.06	0.08	-0.26	0.00
G	-0.44	0.18	-0.14	-0.16	0.12	0.44	-0.21	0.04	-0.39	0.11	-0.35	-0.31	-0.09
H	-0.05	-0.17	-0.14	-0.16	-0.29	-0.28	0.01	-0.50	0.04	-0.31	0.21	0.00	-0.14
I	-0.16	-0.31	0.20	-0.45	-0.06	0.07	0.06	-0.14	-0.39	-0.31	-0.58	-0.48	-0.21
$\phi_{\theta v}$													
A	-0.03	0.13	0.24	0.30	0.24	-0.45	-0.06	-0.12	0.20	0.39	0.30	0.14	0.11
B	0.57	0.82	0.55	0.47	0.44	-0.32	-0.31	-0.12	0.01	0.44	0.26	0.31	0.26
C	0.50	-0.20	-0.28	-0.49	-0.65	-0.13	-0.19	-0.06	-0.26	0.08	-0.27	0.02	-0.22
E	0.29	0.36	-0.05	0.37	-0.02	0.17	0.11	-0.07	0.09	0.16	0.28	0.33	0.17
G	-0.33	-0.20	-0.48	0.25	-0.04	0.00	0.18	-0.05	0.31	0.19	-0.07	-0.55	-0.07
H	-0.52	-0.38	-0.43	-0.03	0.30	0.04	0.11	0.14	-0.10	0.17	-0.40	-0.15	-0.10
I	-0.21	0.03	0.06	-0.13	0.14	-0.08	-0.47	-0.12	0.07	-0.45	-0.38	0.06	-0.12

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Balance of the Transport of Sediment due to Wave Action in Shoaling Water, Surf Zone and on Foreshore*

By

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Abstract

Use being made of an electromagnetic-type current meter, the horizontal velocity of the orbital wave motion was measured in shoaling water, surf zone and on foreshore. The deformation of temporal patterns of orbital wave velocity is discussed with reference to the data from the sampling of suspended sediment, and a simple model on the mechanism of the balance of sediment transport in a cross-section perpendicular to the shoreline is proposed.

1. Deformation of Temporal Pattern of Orbital Wave Velocity in Shoaling Water, Breaker Zone and on Foreshore

Approaching the shore, the swell, affected by bottom friction, undergoes deformation from sinusoidal curve to a shape with sharp crest and flat trough. Simultaneously, the temporal pattern of the horizontal component of orbital wave velocity is deformed; onshore components below crests are sharpened and offshore components below troughs flattened. An example of such deformation, measured with a specially designed electromagnetic-type current meter (NAGATA 1959), is shown in Figure 1. The record A shows a pattern taken at the depth of 10 m, where no serious deformation is seen; the record B taken at the depth of 4 m, showing an increasing trend of deformation; and the record C a typical pattern of deformation just before breaking. All recordings were taken at the level of 20 cm above bottom.

It is highly desirable to measure orbital wave motions simultaneously just before, in, and after breaking and on foreshore in order to know the manner of their progressive deformation. However, breaking does not occur at a definite position. In fact, a breaking wave travels a considering distance in their whole process, which is variable for a crest after another. Therefore, it

* This study was conducted as a part of the beach observation program at Tokaimura under auspices of the Japan Atomic Power Company.

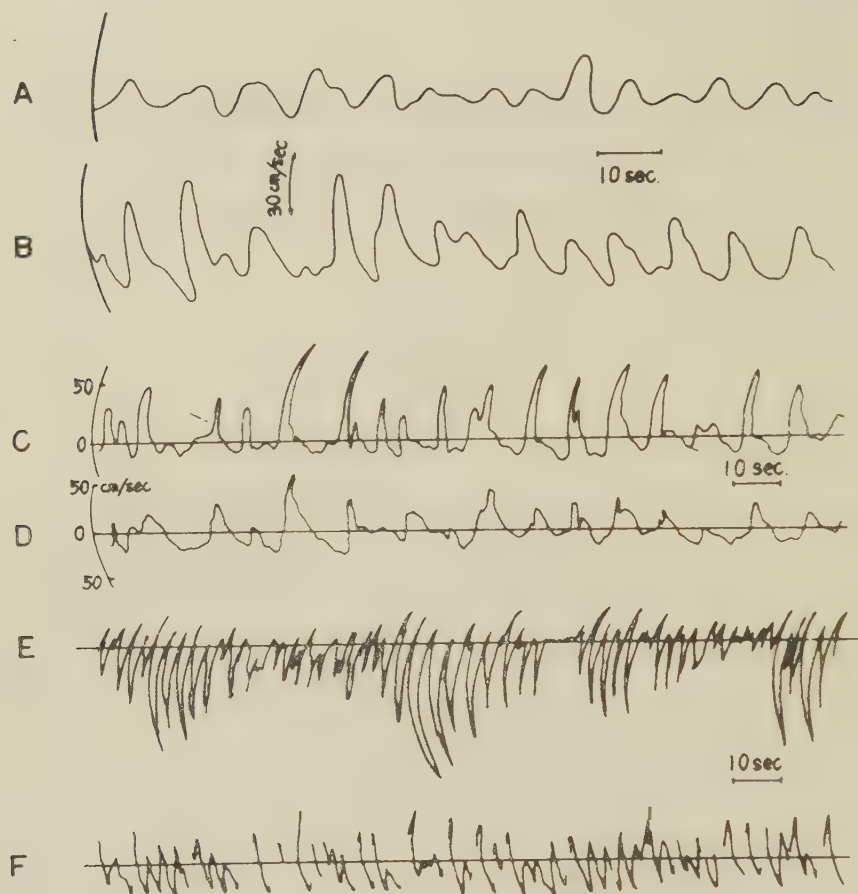


Fig. 1. Deformation of temporal pattern of orbital wave velocity.

A; at the depth of 10 m, B; at the depth of 4 m, C; just before breaking, D; just in breaking, E; just after breaking, i.e. at the lowest level of the foreshore, and F; at the mid-level of the foreshore. Upwards; onshore.

is almost impossible to fix the position of the current meter in relation to each breaker. Besides, high wave conditions render the measurements in the breaking zone too dangerous to risk, while under low wave conditions measurements on the foreshore become practically impossible for want of enough water to keep the current meter submerged throughout the measurement. Consequently, simultaneous measurements in different positions under the same wave condition have not been carried out successfully and the data available here represent the results of the separate measurement under different wave conditions. Discussions to follow are confined to the velocity pattern only.

The records C, D, E and F show typical velocity patterns of orbital wave motion in and near the breaker zone (see Fig. 2 for positions). The records

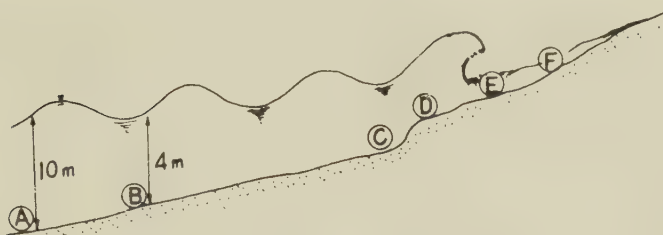


Fig. 2. Positions of measurement.

C and D were taken simultaneously for the breaker heights ranging between 1 to 1.5 m. The record C was taken at the position about 0.5 m offshore away from the foreshore step. The record D was taken at the position about 1 m shorewards from this step. During measurement waves began to break almost above this position D and plunged a little shorewards. The record E was taken on the toe of the foreshore slope corresponding to the plunging position of the breaker for the breaker heights ranging between 2 to 2.5 m. The record F was taken at the mid-level of the foreshore slope for the breaker heights exceeding 4 m. The discontinuities in the record F show that the meter was exposed in air during back wash.

According to the record C, which gives a typical pattern just before breaking, the wave energy is concentrated on the onshore phase of the orbital motion. It is interesting to note that the shift from the onshore to the offshore phase occurs abruptly as seen by a sharp bend in the velocity curve. This pattern, however, appears to be related to the bottom topographies such as steps, for, when the steps are inconspicuous, the velocity curve is more rounded even before breaking.

In the pattern represented in the record D, the shift from the onshore to the offshore component is gradual as compared with the record C, and the influence of back wash from the foreshore is apparently recognized. It is also noted that at this position the peaks of the onshore velocity component tend to be separated into secondary peaks, and that the velocity itself decreases as breaking proceeds.

The record E represents the velocity pattern just after breaking. Upwards from this position which can be regarded as the lowest level of the foreshore, reciprocation of water on the foreshore begins. At this position the velocity pattern greatly differs from the preceding results. The values of the onshore velocity components reach only $1/5$ or $1/6$ those of the subsequent components. However, at the mid-level of the foreshore, the onshore components and subsequent offshore components are almost comparable and the velocity pattern is nearly symmetric as in the record F. This tendency is generally observed on the foreshore. Under the influences of beach topographies such as cusps, or of very high breakers, the details of the velocity pattern are changed, but

it is always observed that at the lower levels of the foreshore the onshore velocity components are much smaller than the subsequent offshore components. This fact can be understood as follows: As the result of breaking, especially in the case of complete plunging breakers, the majority of the incoming wave energy is transformed into turbulent and potential energy giving rise to a sudden elevation of water level immediately shoreward of the breaking position. This elevation of water level has a steep slope on the shoreward front. It is from this mechanism that the onshore velocity component in the upper level of the foreshore is newly generated.

2. Measurements of Onshore and Offshore Transport of Suspended Sediment near the Breaker Zone

The measurement of onshore and offshore transport of suspended sediment was conducted at three positions near the breaker zone, i.e., before, in, and after breaking. The positions of measurements are identical to, C, D and E in Figure 2, where orbital velocity measurements were made. Sufficient accuracy of measurement can not be expected closely to or just inside the breaker zone. Therefore, one of the simplest methods is adopted for the purpose of order estimation. The sand trap consists of two pieces of vinyl pipe (20 cm long and 2 cm I.D.) each equipped with a cotton pouch at one end and a lid at the other. The pipes are fixed to a supporting post in parallel and in opposite directions in order to catch simultaneously the suspended sediment oscillating both onshore and offshore. After they were settled at 20 cm above the bottom and perpendicularly to the shore line at the measuring position, the lids were removed. After they were exposed to ten succeeding wave crests the whole apparatus was pulled out of water. The results of the measurements are shown in the table below.

Table. The results of sampling of suspended sediment.

Sample No.	Position	Direction	Weight	Duration Exposed
1	C	onshore	2.84g	71 sec.
2	C	offshore	7.62	
3	D	onshore	7.71	78
4	D	offshore	27.06	
5	E	onshore	49.43	80
6	E	offshore	15.63	

The summed values of suspended sediment caught in both directions at each position C, D and E have the ratios 1:4:7. It follows that the volume of suspended sediment transport is largest at the position just onshore of breaking and that it decreases rapidly toward offshore.

At the lowest level of the foreshore located just onshore of the breaker zone the volume of the onshore transport is three times that of the offshore transport. On the contrary, in the breaker zone and just offshore of the breaker zone the volume of the onshore transport is only 1/3 of the offshore transport. Even though allowances should be made of the characteristics of the sand trap to velocity change, these differences between the volumes of the onshore and offshore transport are significant enough. Therefore, the observation shows that the sediments set into suspension by breaking waves are carried in both directions, onshore and offshore.

In Figure 3 the cumulative distributions of sand size are shown for each sample. The characteristics of each set of sample are clear. At the lowest

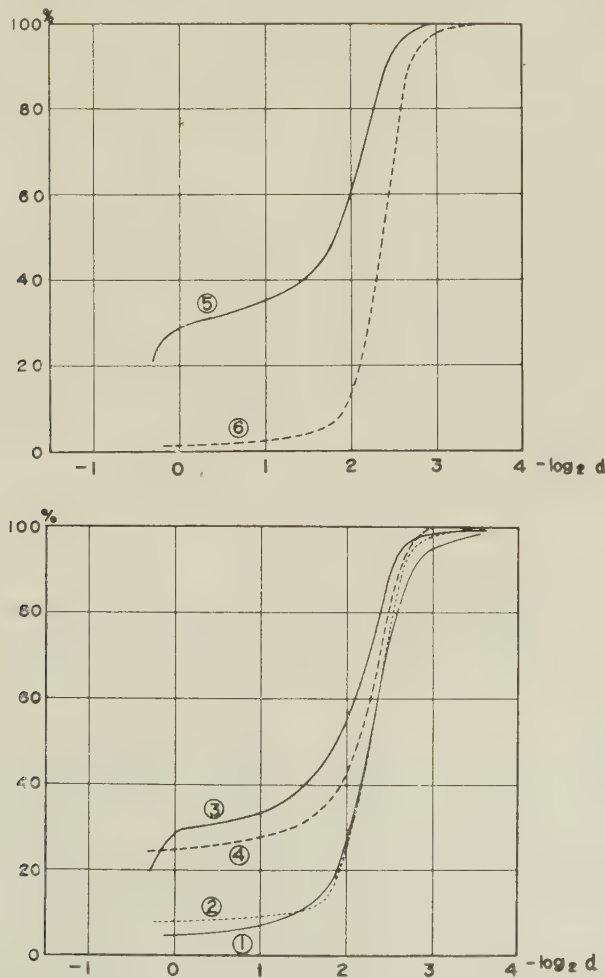


Fig. 3. Accumulative distributions of sand size of samples from sand trap measurement.

level of the foreshore a sharp difference exists between the size distributions of the sediments transported onshore and offshore. The onshore sediment contains a great deal of small pebbles, while the offshore sediment the ordinary sizes only. On the contrary, the samples from in and just offshore of the breaker zone do not show any significant difference between the onshore and offshore sediments. The samples from the breaker zone show the similar distribution to the onshore sediment from the lowest level of the foreshore, while that from just offshore of the breaker zone is rather similar to the offshore sediment from the lowest level of the foreshore. It is presumed that these characteristics are related to the conditions of turbulence at these positions, i. e., in the breaker zone, between the bottom and the 20 cm level at lowest, turbulence is strong enough to lift small pebbles regardless of the direction of the orbital velocity components.

In view of these facts it seems to be reasonable that the volume of the offshore transport in the breaker zone is larger than that of the lowest level of the foreshore as shown in the preceding table though the summed transport is much smaller.

3. Balance of Water Mass and Sediment on Foreshore

The velocity pattern caused by the reciprocating water mass on the foreshore has already been discussed in Chapter 1. The velocity pattern is asymmetric at the lower level of the foreshore and symmetric at the mid-level of the foreshore. This fact will be explained by evaluating the balance of water mass on the foreshore. We can now assume that the total volume of water carried onshore through a considered section over an adequate length of time is just equal to that of water returning offshore, and hence.

$$\int_T H V dt = \int_{T'} H' V' dt, \quad (1)$$

where H and V are, respectively, the water level and the velocity of water particle at the considered section during the onshore motion T ; and H' and V' , respectively, are the water level and the velocity of water particle during the offshore motion T' .

Assuming that these durations are approximately equal, and considering $H \gg H'$ at the lower level of foreshore, we have a relationship $V \ll V'$. At the upper level of the foreshore we may assume $H \approx H'$ and have $V \approx V'$.

The foreshore topography is most variable in the cross-section perpendicular to the shoreline. However, the time scale is assumed in the present discussion in such a manner that the foreshore topography is regarded to be in the state of equilibrium. Therefore, the transport of sediment must be balanced in the first approximation. Denoting the volume of suspended sediment concentration in onshore and offshore motions with Q and Q' , respectively,

we can express the volumes of onshore and offshore transport of suspended sediment, respectively, as follows;

$$\int_T QHVdt, \quad \int_{T'} Q'H'V'dt.$$

At the lower level of the foreshore it is evident that $Q \gg Q'$. And considering the equation (1), we have the following relationship;

$$\int_T QHVdt \gg \int_{T'} Q'H'V'dt. \quad (2)$$

Therefore, the residual sediment must be carried back offshore along the bottom by saltation and creeping. It is supposed that characteristic velocity pattern $V \ll V'$ plays an important role in this process. At the upper level of the foreshore it is reasonable to assume that $Q \approx Q'$, and also considering the relationships $V \approx V'$ and $H \approx H'$, it is self-evident that there exists the balance of sediment transport.

4. Balance of Sediment in and just Offshore of Breaker Zone

As the characteristics of these regions, we know the following facts from the observations; (1) the velocity pattern in orbital wave motion undergoes severe deformation, and the onshore components are much larger than the offshore components; (2) the onshore transport of suspended sediment is only 1/3 the offshore transport; and (3) the concentration of suspended sediment decreases toward offshore and the size distribution does not show a significant difference between the onshore and offshore sediments.

In discussing the balance of sediment at the lower level of the foreshore we supposed that the unbalance in the velocity pattern results in the transport of sediment along the bottom by saltation and creeping in the direction of larger velocity component. And it seems to be reasonable to deduce from the fact (1) that the transport of sediment along the bottom is directed onshore in and just offshore of the breaker zone. In view of the sediment balance this deduction agrees with the fact (2) as seen from the result of the measurement.

The fact (3) explains the fact (2). That is to say, the existence of the sudden decrease towards the offshore in concentration of suspended sediment would cause the offshore transport of suspended sediment by diffusive action. In other words, the water mass moving offshore has a relatively larger amount of suspended sediment than the water moving onshore, for the turbulent energy is decreased toward offshore. Therefore, the net transport of suspended sediment is directed offshore. In conclusion, sediment is transported onshore along the bottom due to unbalanced velocity pattern of the wave motion and transported offshore in suspension due to diffusive action.

5. Effect of Offshore Current on Sediment

From the assumption that the velocity pattern is related to the direction of transport of sediment along the bottom, the mechanism of the balance of sediment transport at the further offshore positions in the cross-section will be deduced. Transport of sediment along the bottom is apparently directed shorewards at every position ranging from the depth where the velocity pattern of orbital wave motion undergoes effective deformation, to the breaker zone. Therefore, resultant sediment transport along the bottom must be directed offshore. But we cannot suppose the existence of concentration gradient of suspended sediment in this region except for the immediate neighbourhood of the breaker zone. Consequently, it is thought that the offshore current system plays an important role in the balance of sediment.

In the above discussion, we dealt with the case of the narrow breaker zone or a breaker line. In the case of the broader zone of breaking, the velocity pattern of orbital wave motion will be similar to that shown on the record D, and transport of sediment along the bottom due to the wave action will be shorewards. However, since an effective concentration gradient of suspended sediment may not be expected, the role of the undertow near the bottom must be taken into account for the balance of sediment.

At the coast of Niigata the author observed during small wave conditions coexistence of a strong undercurrent and a strong offshore surface current in and offshore of the breaker zone respectively (Nagata 1959). This seems to support the conclusions derived in this chapter.

6. Discussion of Assumption

It is necessary to discuss the assumptions used in the previous chapters. In assuming that only the unbalance in the velocity pattern of orbital wave motion is responsible for the direction of transport along the bottom, we neglected the effect of the difference of the durations between onshore and offshore motion. Mass transport of water must be balanced everywhere, so the equation (1) is applicable to the positions in and just offshore of the breaker zone. It is apparent that $V \gg V'$, and also $H \gg H'$ in these regions. So, the relationship $T \ll T'$ is required from the equation (1). Therefore, the difference of durations plays an important role in the balance of water mass.

Denoting the onshore and offshore transport along the bottom with $T_b(V)$ and $T_b'(V')$, respectively, we have the following equation for the balance of sediment,

$$\int_T QH V dt + \int_T T_b(V) dt = \int_{T'} Q' H' V' dt + \int_{T'} T_b'(V') dt. \quad (3)$$

As the relationship $Q \ll Q'$ is apparent from the concentration gradient of suspended sediment, we have the following relationship,

$$\int_T QHVdt \ll \int_{T'} Q'H'V'dt. \quad (4)$$

Therefore,

$$\int_T T_b(V)dt \gg \int_{T'} T_b(V')dt. \quad (5)$$

If $T_b(V)$ increases linearly with the increase of current velocity as represented by,

$$T_b(V) = kV, \quad (6)$$

the following relationship must hold

$$\int_T T_b(V)dt (=k \int_T Vdt) \ll \int_{T'} T_b(V')dt (=k \int_{T'} V'dt) \quad (7)$$

since,

$$\int_T Vdt \ll \int_{T'} V'dt. \quad (8)$$

from the relationship $H \gg H'$ and the equation (1). Therefore, the transport of sediment along the bottom must increase more rapidly with the increase of current velocity, as represented by,

$$T_b(V) = kV^n \quad (n \gg 1).$$

This is the assumption used in the previous discussion. And the effect of the difference of durations between the onshore and offshore motion was neglected in the previous discussion on the balance of sediment.

7. Conclusions

From the observed data, a simple but self-consistent model on the mechanism of the balance of sediment in the cross-section perpendicular to the shoreline is obtained in equilibrium state. The discussion dealt only with the balanced state as the first approximation. Although the author refrained from saying anything definite about the erosion and accretion, it shows one of the approaches to the problem on the mechanism of erosion and accretion of a beach for a future investigation and will also serve as a good reference for the plans of beach observations and model experiments.

The author wishes to express his thanks to Professor Koji HIDAHA, University of Tokyo, for his guidance and encouragements. His thanks are also due to Professor Masashi HOMMA, Assistant Professor Kozo YOSHIDA and Assistant Professor Kiyoshi HORIKAWA, University of Tokyo, for their helpful suggestions, and staff engineers of the Japan Atomic Power Company for their assistance during the course of this study.

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An Analysis of the Mechanism of the Cold and Warm Water Masses in the Seas adjacent to Japan

By

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Abstract

In the present paper the large cold water mass in the offing of the Enshunada and the warm water mass in the Japan Sea are analyzed. From this analysis it is seen that the appearance of these water masses has a close relation to the bottom topography.

1. Introduction

We have many examples of the appearance of the cold and warm water masses in the seas adjacent to Japan. The large cold water mass in the offing of the Enshunada is one of the most characteristic phenomena in the Kuroshio, and we can not recognize the persistence of the cold water mass in any ocean currents other than the Kuroshio, for example in the Gulf Stream of the Atlantic Ocean corresponding to the Kuroshio of the Pacific Ocean. Besides, we can find that the warm water mass in the Japan Sea occupies almost the same place constantly. These warm and cold water masses may be said to have a characteristic of steady stay. It is natural that oceanographers in our country are strongly interested in these phenomena. The author has collected the data about the large cold water mass in the Pacific Ocean and those about the warm water mass in the Japan Sea, and analyzed the mechanism of the cold water mass appearance in the Enshunada and of the warm water mass in the Japan Sea, in order to search for methods of forecasting the appearance of these masses.

The present analysis has hardly established any theory about the appearance of cold water mass, but it may be important to pursue the data conscientiously and to study the Kuroshio and the Tsushima current. (The positions of islands and capes in this paper are indicated in Fig. 1). In this paper the analysis of the large cold water mass is made at first, and then the warm water mass is studied.

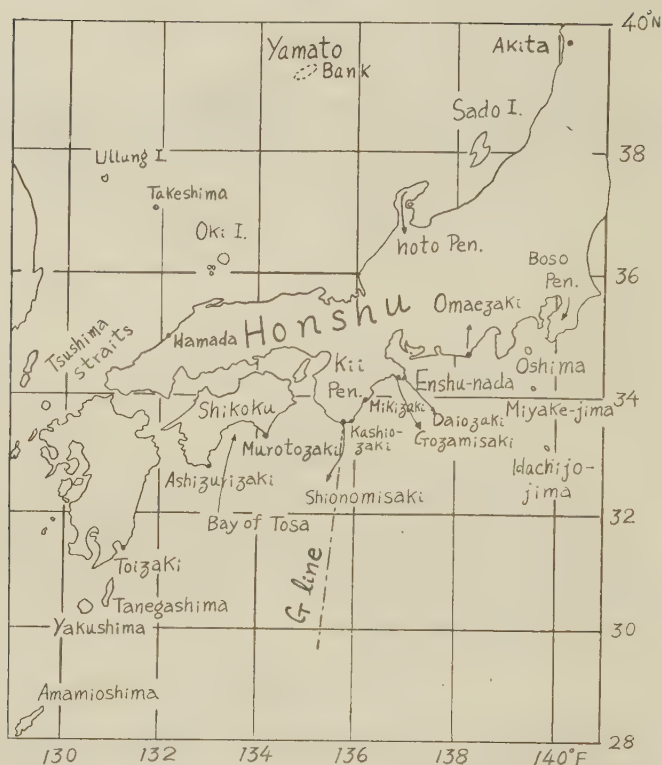


Fig. 1. Positions of Island, Cape and Bay.

2. The appearance of the large cold water mass in 1959

In March 1959 the Kuroshio streamed very near Shikoku from west to east, but in May the offshore receding of the Kuroshio axis was found by oceanographical observations on board the Shumpu maru, Kobe Marine Observatory.¹⁾

This offshore receding of the Kuroshio axis is very curious, never seen before (Fig. 2). In company with this offshore receding, a low water temperature region is formed between Shikoku and the Kuroshio axis.

In June the flow pattern of the Kuroshio presents an abnormal mode (Fig. 3). Namely, the meandering of the Kuroshio seems to go eastward, and the low water temperature region shifts from south of Shikoku to south of the Kii peninsula.²⁾ At last the variation of the Kuroshio flow pattern attains a typical large cold water mass pattern which is already known to us, and the cold water mass occupies the offing of the Enshunada. The variations of the Kuroshio axis are shown in Fig. 4. Summarizing these variations, we can probably assume the abnormal state of the Kuroshio off Shikoku to be a pre-sage of the appearance of the large cold water mass in the Enshunada.

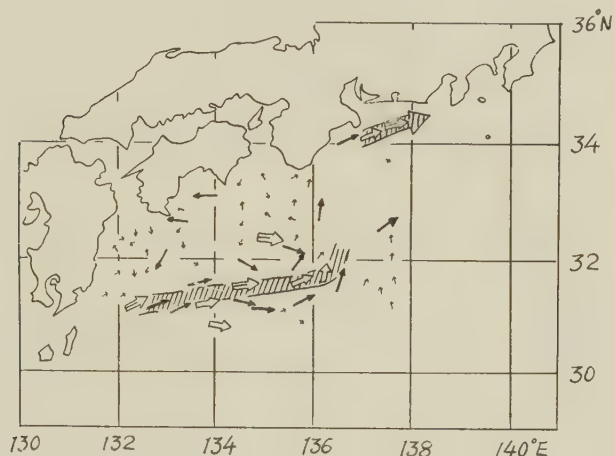


Fig. 2. Surface current by GEK May 1959 (Shumpu maru).

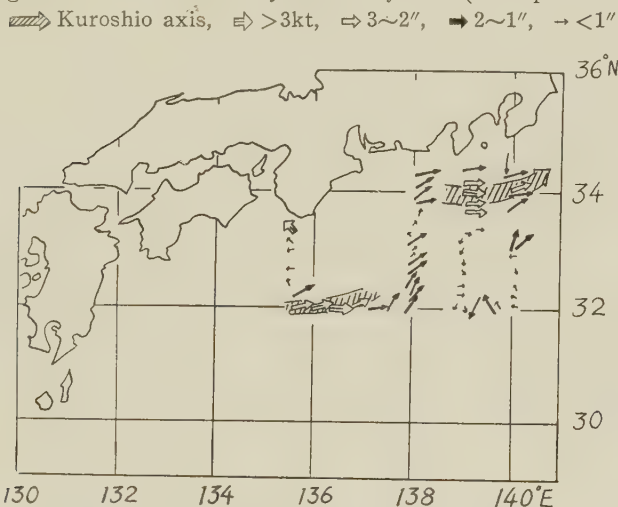


Fig. 3. Surface current by GEK, June 1959 (Hydrographic Department).

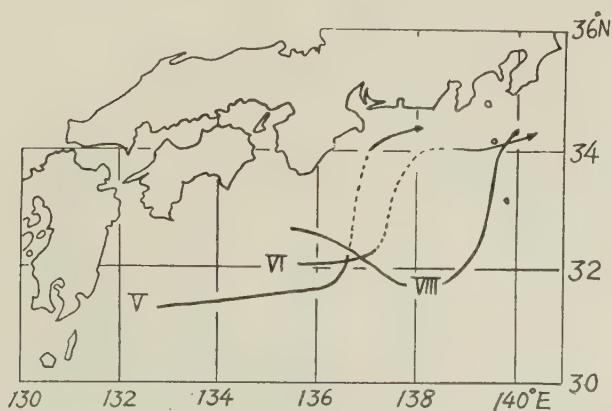


Fig. 4. Variations of the Kuroshio axis (V: May VI: June VIII: Aug.).

We have already two examples of the large cold water mass appearance, i.e. from 1934 to 1943 and from 1954 to 1955. So it is necessary to know from the observational data whether the presage of the cold water mass appearance is shown by the abnormal state of the Kuroshio off Shikoku or not.

3. The large cold water mass from 1934 to 1943

We have some rather accurate oceanographical data since 1927. For analyzing the mechanism of the cold water mass, the data before 1927 can not be effectively utilized, because serial oceanographical observations were not sufficiently carried out at that time and the data are very scanty.

At first the author has minutely studied the reports on the ocean current in the Hydrographic Bulletin³⁾ published by the Japanese Hydrographic Department and in the General State of the Seas in the Oceanographical Investigation⁴⁾ published by the Imperial Fisheries Experimental Station, with a view of finding an abnormal ocean current and a presage of the large cold water mass. These ocean current reports are summarized in Table 1.

From 1927 to 1933 the Kuroshio off Shikoku is noticed to stream normally eastward and almost along the Pacific coast of Honshu. Fig. 5a shows the summarized ocean state mentioned above. The flow pattern is shown by the distribution of isotherms at 100-m layer (Fig. 5). The most characteristic features seen in Fig. 5a are the straight streaming of the Kuroshio and the existence of the Kuroshio axis near Honshu. Such a Kuroshio flow pattern is generally said to be normal. In January 1933 the cruiser Chokai of the Japanese Imperial Navy reported on an abnormal state of the Kuroshio near Shikoku for the first time. Although the Kuroshio flow pattern returned to

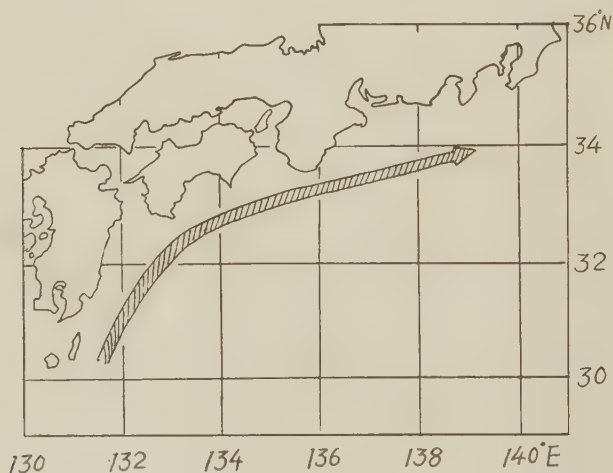


Fig. 5-a. Sketch map of the Kuroshio (1927-1933).

Table 1. The Condition of the Kuroshio from ocean current and water temperature reports.

year	Month	Hydrographic Bulletin		Oceanographical Investigation		
		Ocean current	observer (Ship name)	Year	Month	General State of Seas
1927	Feb.	South of Shionomisaki E 2~3kts	Hydrographic Department	1931	July~Dec.	50 nautical miles south of Gozamisaki, Kuroshio
	Aug.	The north margine of the Kuroshio near Shionomisaki	Seito			
	Nov.	Shionomisaki~Murotozaki E, 1kt near Ashizurizaki current weak	Naka	1932	Aug. Oct. Feb. Apr.	Kuroshio near the Bay of Tosa Kuroshio 40~70 N. M. south of Gozamisaki.
1928	July~Aug.	Kuroshio near Tosa Bay	Hydro. Dep.	1933	June Oct. Mar.	Kuroshio near the Bay of Tosa Kuroshio 20~70 N. M. south of Gozamisaki
1929	Sept.	Kuroshio near Tosa Bay	"		Apr.	Feb.~Mar. Kuroshio offshore south of Ashizurizaki
1930	Jan.	Kuroshio path normal			May Nov.	Nov. Kuroshio 50 N. M. south of Ashizurizaki
1931		31°10'N 30°57'N } 131°35'E ~132°00'E }	Chokai	1934	Feb.	Kuroshio recedes south of Shikoku (80 N. M.)
1932		30°57'N 31°18'N } 132°00'E ~132°18'E }	"	"	May	"
"	Feb.	32°35.5'N } 134°21' E 62° 2.8kts	"	"	Aug.	large cold water mass appears (Aug.~Dec.)
"	Mar.	32°18'N 31°40.5'N } 134°08'E ~134°11' E } 355° 0.7kts	"			
"	"	Ashizurizaki~Murotozaki W. 0.4kts	Tairyu-maru	1935	Feb.	Kuroshio recedes south of Shikoku
"	"	Toizaki~Ashizurizaki ocean current very weak	Ganges-maru			
"	May	South of Kii pen. Ocean current not seen	Heimei-maru			
"	Dec.	Toizaki~ 33°04'N 134°57'E } 247° 0.7kts	Gyoko-maru		June	high water temperature south of Gozamisaki (highest since 1919)
"	"	Toizaki~Shionomisaki Ocean current not seen	Osaka-maru		Oct. Nov.	large cold water mass developes high water temperature south of Gozamisaki
1934	May	Murotozaki~Shionomisaki W 0.6kts	Tomitura-maru	1936	Feb.	low water temperature south of Shikoku
"	Sept.	"	Amazon-maru			
1935	May	Ashizurizaki~Shionomisaki counter current	Hydro. Dep. Kojun-maru			Kuroshio recedes south of Shikoku (90N. M.)
		South off Shionomisaki, Kuroshio northward			Jan.~Oct. Sept.~Oct.	large cold water mass extreme high water temperature south of Gozamisaki
1936	Feb.	Tanegashima~ 32°45'N ~Shiono- 134°54'E } misaki 245° 1kt, 249°1.9kts	Senzan-maru	1937	Feb.	Kuroshio approaches the Bay of Tosa, large cold water mass off Shionomisaki
"	"	31°27'N 33°20'N } 132°15'E ~135°43'E } 239° 1kt	Aso-maru			
"	"	Shionomisaki~Mikizaki 60° 3~4kts	Kanda-maru		Apr. May June	large cold water mass off Shionomisaki, but high water temperature south of Gozamisaki
"	"	Daiozaki 55° 3kts	Tatsuta-maru			
"	Sept.	Omaezaki 65° 2.4kts Shionomisaki 85° 2kts Daiozaki 68° 4kts Kashinozaki 58° 4kts	Chicago-maru		July Aug. Oct. Nov.	developed large cold water mass Through a year cold water mass developed
"	Oct.	32°31'N 133°52'E } 111° 0.6kts	France-maru	1938		
"	"	West of Shionomisaki 275° 0.6kts	Isuzu-maru		June	Kuroshio approaches the Bay of Tosa
"	"	Shionomisaki~Omaezaki 70° 2.3kts	Sanramon-maru	1939	May~Dec.	Cold water mass developes, but in Apr. May high water temperature south of Gozamisaki
1937	Feb.	off Ashizurizaki Kuroshio normal, off Shionomisaki Kuroshio Southward	Shunsho-maru	1940	Apr.~Aug. Dec.	Cold water mass developes Through a year Cold water mass developes
1938	Jan.	33°35'N 130°45'E } 342° 0.6kts	Kaijo-maru	1941		
"	July	Shionomisaki~Kashinozaki 234° 1.9kts	Daihozan-maru			
"	Sept.	30°40'N 32°05'N } 136°56'E 135°11'E } 127° 2kts	Shintoku-maru			
1940	Nov.	32°41'N ~Kashinozaki 307° 0.35kts 136°56'E }	Kyushu-maru			

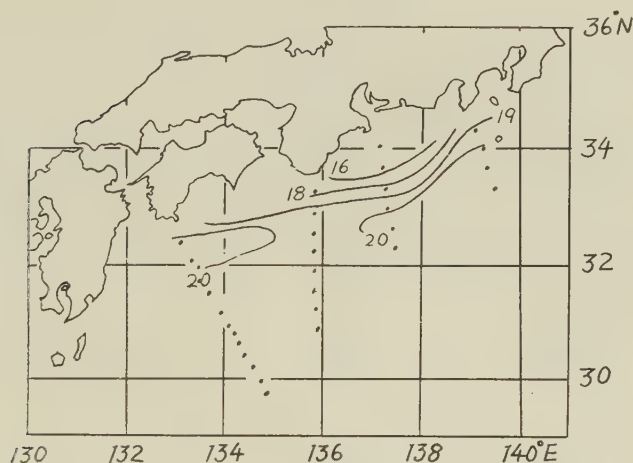


Fig. 5-b. Typical isotherms on 100 m layer from
1927 to 1933 (June, 1932).

be normal thereafter, observations of the westard counter current in the Kuroshio region off Shikoku were often reported (see Table 1). Oceanographical observations were carried out by the Hydrographic Department in April 1933, concerning the synoptic hydrographic condition. In the map prepared from these observations the Kuroshio south of Shikoku shifts rather southward and the cyclonic eddy exists south of Shikoku. (Fig. 6) The mean location of the Kuroshio axis south of Ashizurizaki is expressed by the following formula based on the oceanographical observations from 1956 to 1958 which are carried out accurately by new and improved instruments.

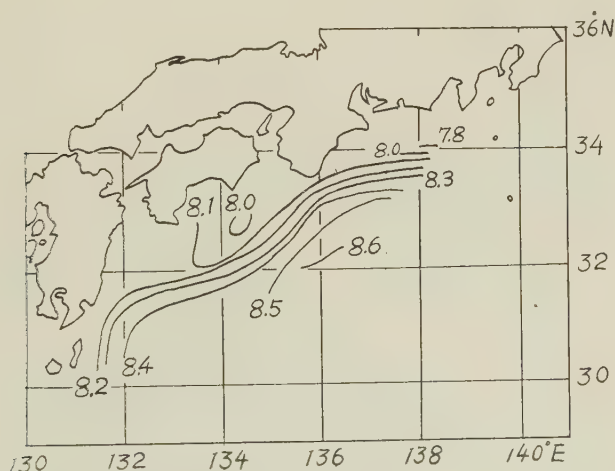


Fig. 6. Dynamical topography, April 1933
(0-800db, +770) (Hydrographical Department)

$$L=40\pm 10 \text{ nautical miles}$$

L : (distance in the southsoutheast direction from Ashizurizaki to the Kuroshio axis)

Now we can estimate that if the location of the Kuroshio axis exists 50 nautical miles south of Ashizurizaki, it is farther than the mean location. Oceanographical observations off Gozamisaki in February 1934 seems to suggest that the large cold water mass may appear off the Kii peninsula, but the observational data are too scanty to assert the occurrence of the cold water mass. In August 1934 the cold water mass distinctly appeared off the Kii peninsula⁵⁾ (Fig. 7).

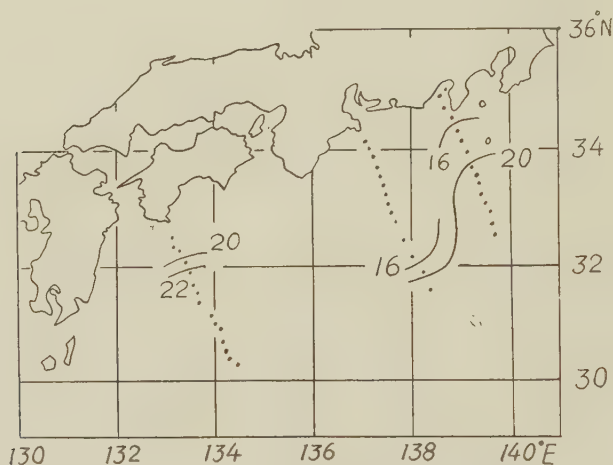


Fig. 7. Isotherms on 100 m. Aug. 1934

According to these data, it is seen that the abnormal state off Shikoku may occur several months (from 2 or 3 months to about a year) before the appearance of cold water mass off the Kii peninsula. The offshore receding of the Kuroshio axis in 1933 and 1934 is also ascertained in the offing south of Shikoku by the report on General State of Seas in the Oceanographical Investigation (see Table 1). Namely, this phenomenon is similar to the offshore receding of the Kuroshio axis south of Shikoku in 1959, but by a study in detail, the offshore receding of the Kuroshio axis south of Shikoku in 1933 is not so remarkable as in May 1959. The difference in the both cases is not here discussed.

In the early stage of the cold water mass, its location varies with seasons. An example of the variations of the location is shown in Fig. 8. In June and November 1935, the water temperature south of Gozamisaki is higher than the mean, but there is always the cold water mass in February, March, August and October. Besides, in June and November we have some data for assuming a region of low water temperatures south of the Kii peninsula. From these

phenomena it is seen that the cold water mass moves sometimes from west to east, and sometimes from east to west. When the cold water mass is transferred from east to west, the northward or northeastward Kuroshio is observed near Gozamisaki (see Table 1, February and September, 1936). This northward strong Kuroshio naturally necessitates an isothermal distribution at the 100-m layer, which shows low temperatures south of the Kii peninsula and high temperatures south of Gozamisaki (Fig. 8). Such eastward movement of the cold water mass is seen until the end of 1937. However, in 1938 the cold water seems to occupy its settled location⁶⁾. The location of cold water mass from 1938 to 1941 are shown in Fig. 9, from which we can deduce that the developed large cold water mass occupies a certain space constantly and the

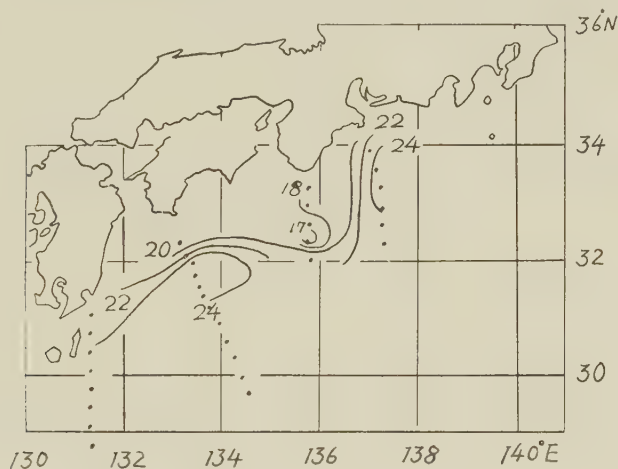


Fig. 8. Isotherms on 100 m, Oct. 1936.

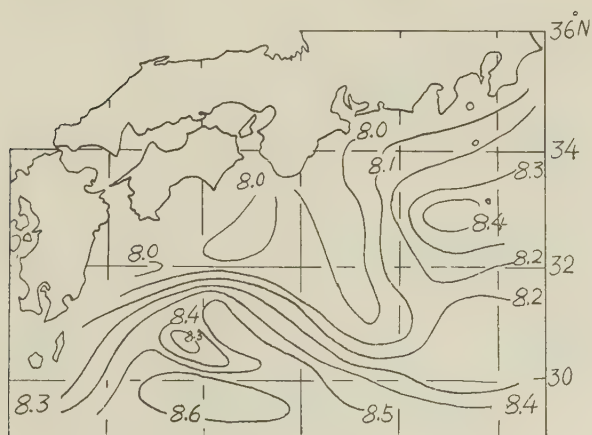


Fig. 9-1. Dynamical topography, Jan. 1937.
(0-800db +770) (Hydrographic Department)

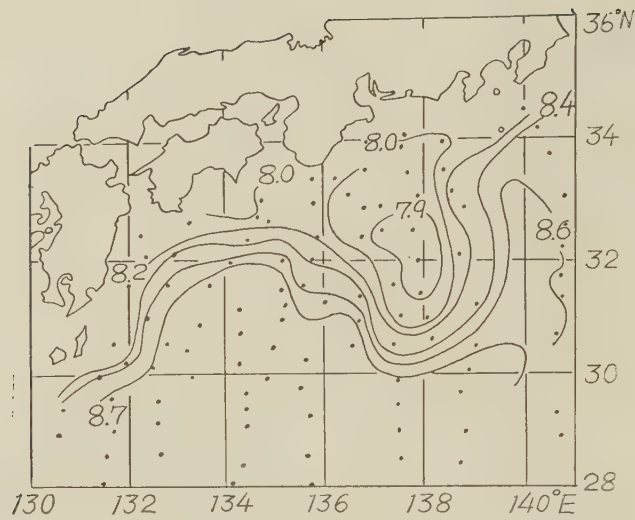


Fig. 9-2. Dyn. top. Aug. 1938.
(0-800db +770) (Hydr. Dep.)

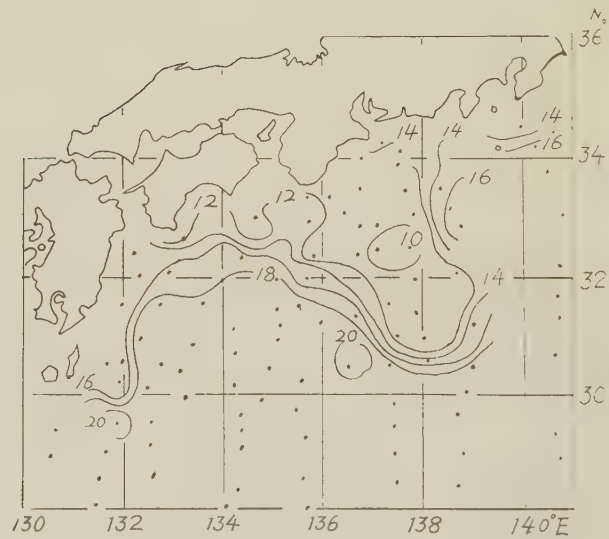


Fig. 9-3. Isotherms on 200 m, Aug. 1938.

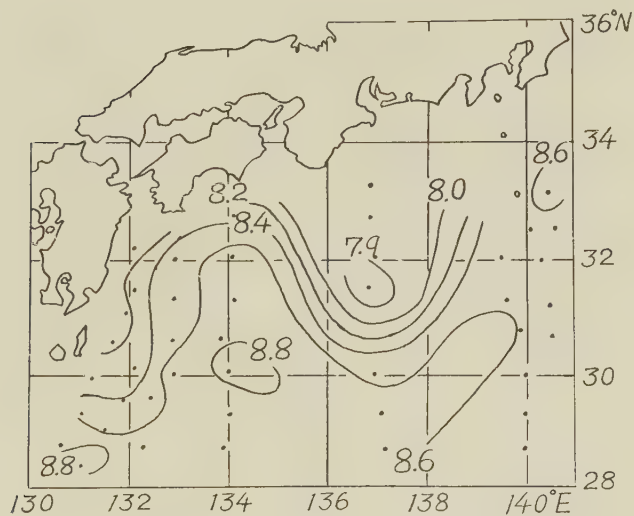


Fig. 9-4. Dyn. top. July, 1939.
(0-800db, +770) (Hydro. Dep.)

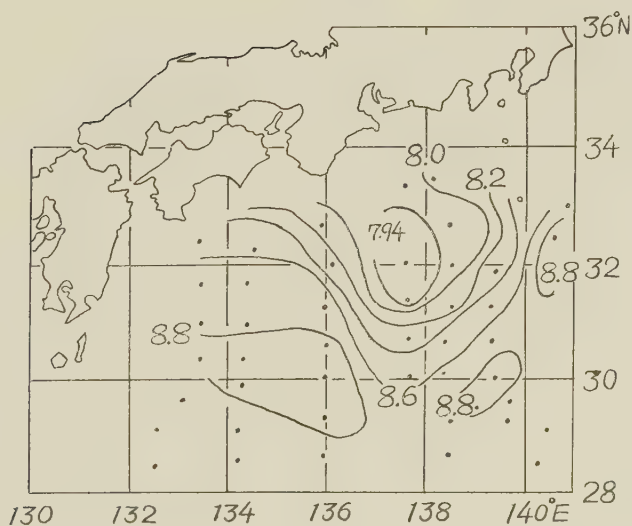


Fig. 9-5, Dyn. top., Aug. 1939.

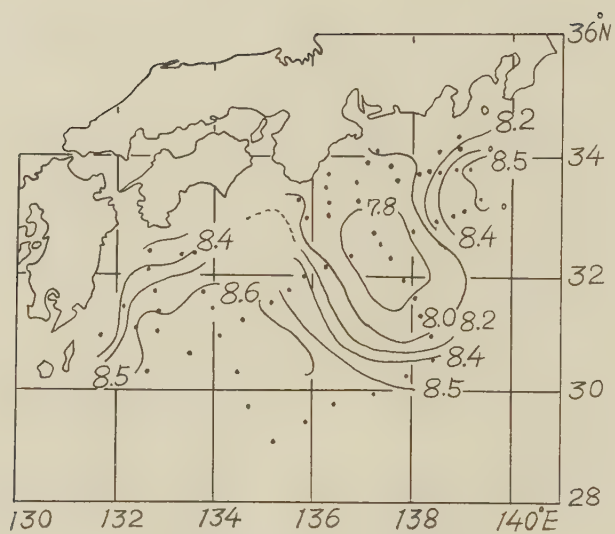


Fig. 9-6. Dyn. top., Nov.~Dec. 1939.

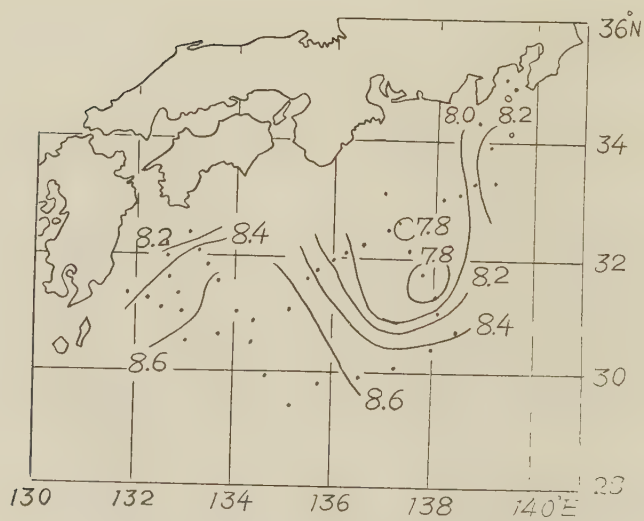


Fig. 9-7. Dyt. top. Jan.~Feb., 1940.

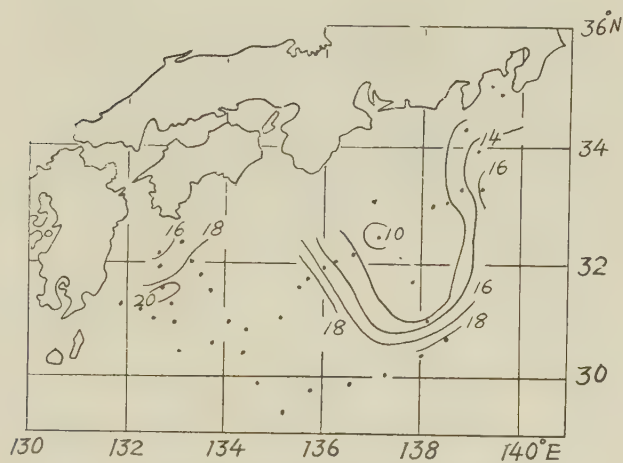


Fig. 9-8. Isotherms on 200 m. Jan.~Feb., 1940.

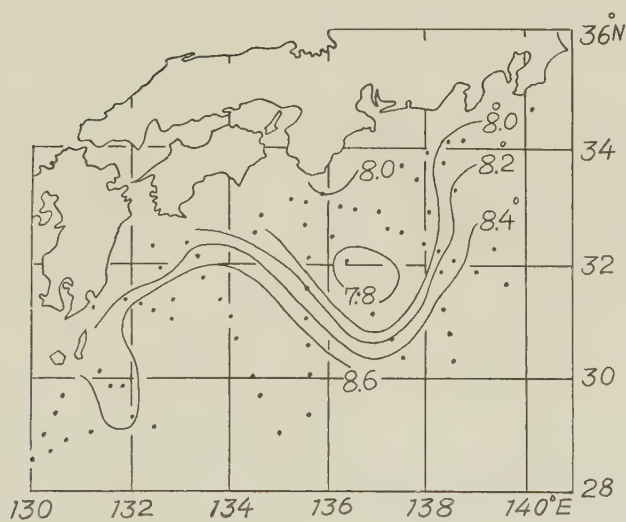


Fig. 9-9. Dyn. top., June 1940.

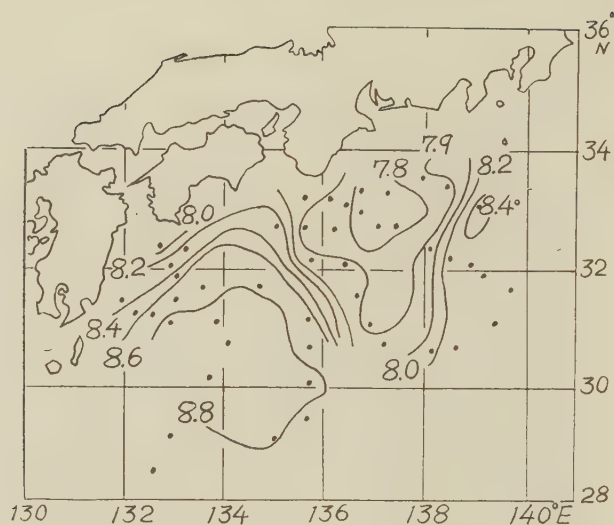


Fig. 9-10. Dyn. top., Aug., 1940.

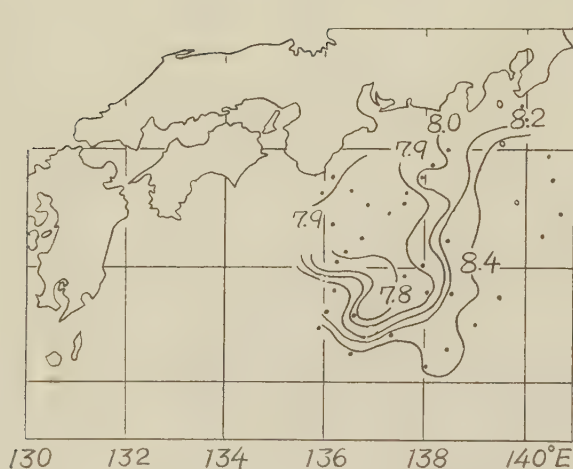


Fig. 9-11. Dyn. top., Nov., 1940.

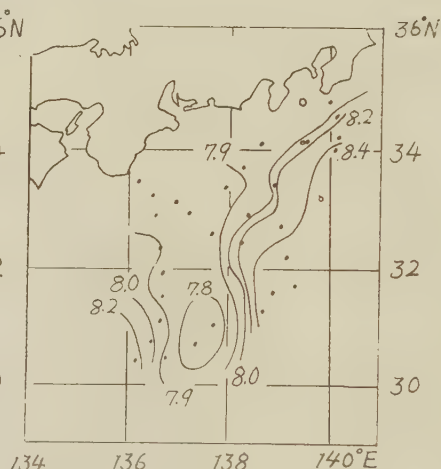


Fig. 9-12. Dyn. top., Jan., 1941.

Kuroshio approaches the shore near Shikoku and then recedes off the Kii peninsula or in the offing of the Enshunada. Through 1941 the large cold water mass exists in the same space. Thereafter during the 2nd World War, the oceanographical observations could not be carried out sufficiently. By referring to the data observed by the Japanese Imperial Navy and to the reports from fishing boats, however it may be said that the cold water mass existed surely in April 1943. And then in December 1943 the large cold water mass may be reduced in its scale⁶⁾ (Fig. 10). After the War this cold water mass disappeared.

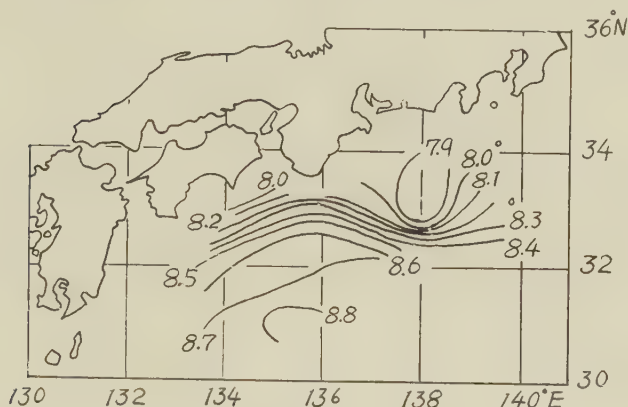


Fig. 10. Dynamical topography, Dec., 1943.
(0-800db, +770) (Hydro. Dep.)

4. The large cold water mass of 1954

After the 2nd World War we did not see the large cold water mass developed in the offing of the Enshunada and could not recognize the large cold water mass until 1954. But in May 1954 it was reported by the Shumpu-maru, Kobe Marine Observatory^{7),8)}, and by the Atsumi, an ocean weather ship, respectively, that the large cold water mass appeared. This cold water mass was existing through 1954 and 1955.

Now the author would like to define the large cold water mass in the following. Isotherms of 12°-14°C at the 200-m layer reached 31°N in the offing of the Enshunada and isobars of closed type is in existence at about 31°N. Such are the characteristics of the large cold water mass. According to this definition it may be said that the flow patterns of the Kuroshio from 1946 to 1952 are normal. (In Fig. 11 are shown current charts of dynamical topography and isothermal maps for the period from 1946 to 1958.) Next, investigation is made on the presence of a presage of cold water mass. In order to clarify the approaching and receding of the Kuroshio axis, the descriptions by the Kochi Fisheries Experimental Station⁹⁾ are reproduced on Table 2. According this table, it is known that in November and December 1953 the Kuroshio axis receded to the south off Shikoku. On examining these phenomena stated above, it might be correct to conclude that the offshore receding of the Kuroshio axis occurs several months before the appearance of the large cold water mass in the Enshunada.

We have discussed the large cold water mass, but in the normal pattern of the Kuroshio we can see frequently the small cold water mass. In July 1946 a small cold water mass is formed in the Enshunada, and the Kuroshio flows rather south of the Enshunada. In June 1947 the cold water mass is

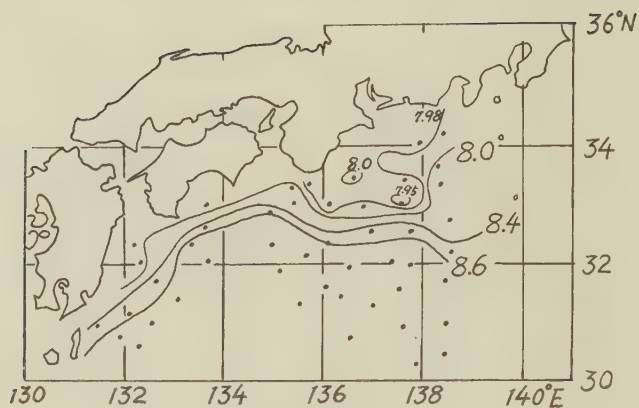


Fig. 11-1. Dynamical topography. July~Aug., 1946.
(0-800db, +770) (Hydro. Dep.)

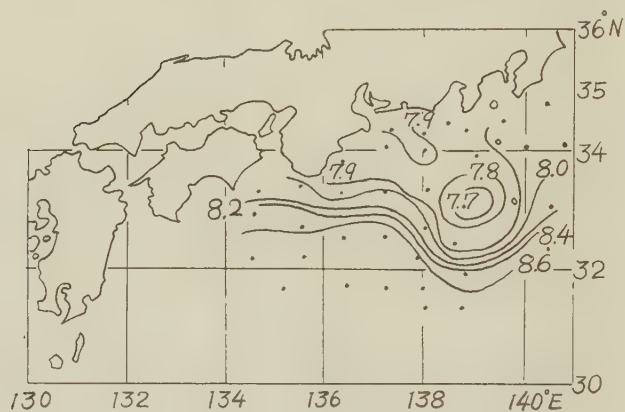


Fig. 11-2. Dynamical topography, June 1947.
(0-800db, +770) (Hydro. Dep.)

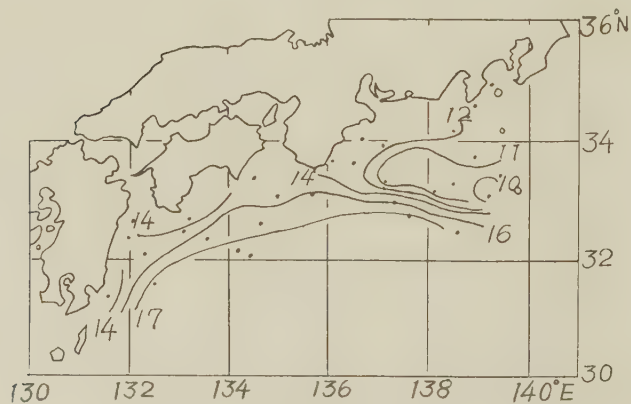


Fig. 11-3. Isotherms on 200 m, Apr. 1948.

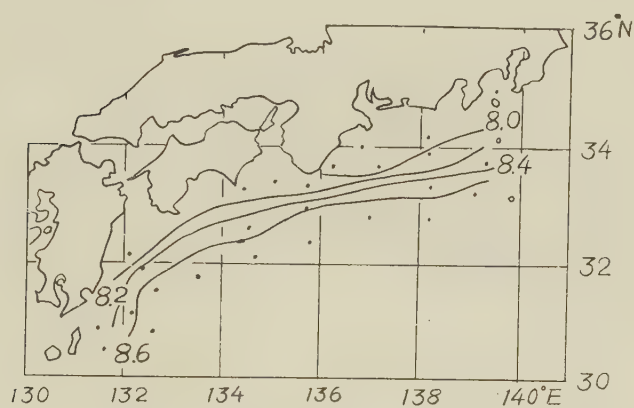


Fig. 11-4. Dyn. top. Nov.~Dec., 1950.

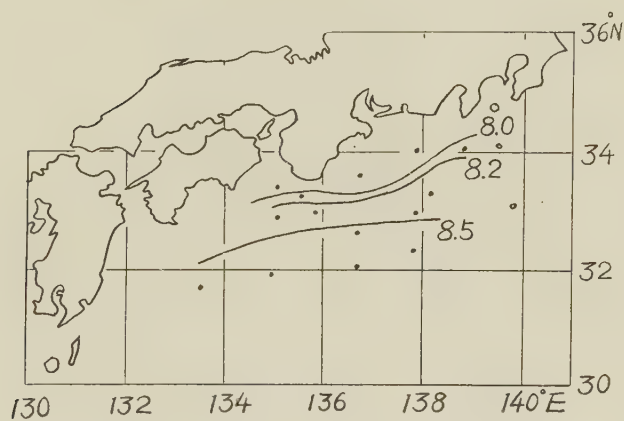


Fig. 11-5. Dyn. top., Feb.~Mar. 1951.

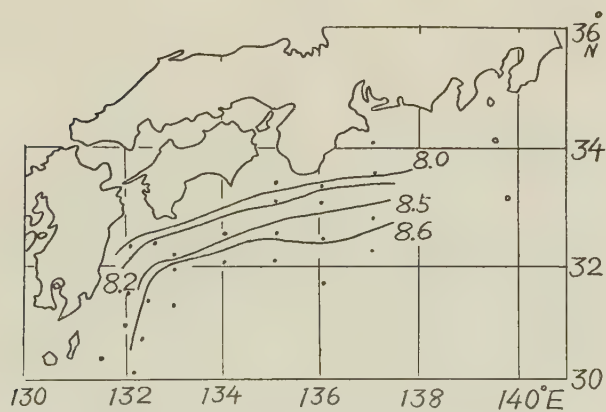


Fig. 11-6. Dyn. top. May, 1951.

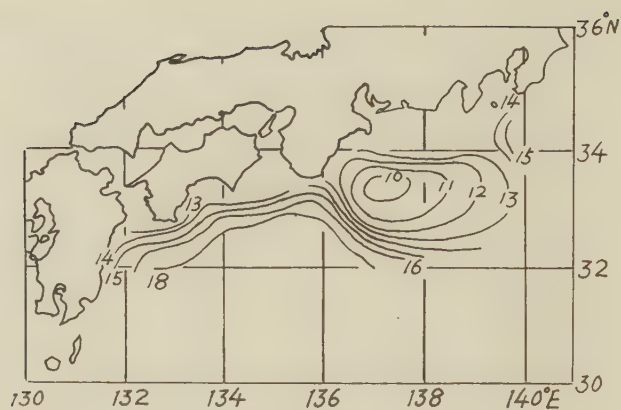


Fig. 11-7. Isotherms on 200 m, Feb.~Mar., 1952.

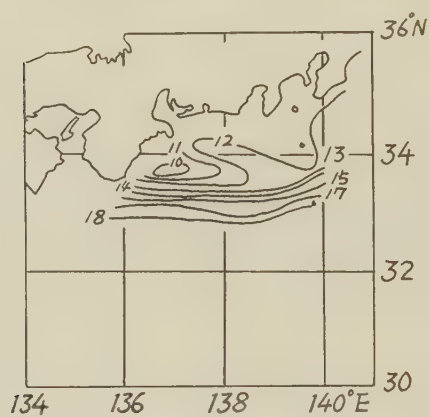


Fig. 11-8. Isotherms on 200 m, May, 1952.

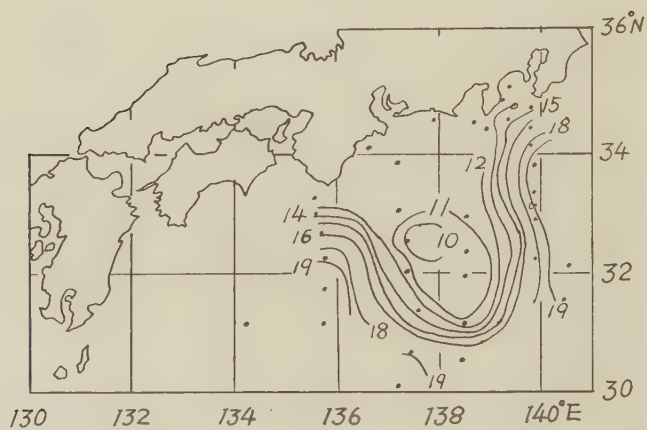


Fig. 11-9. Isotherms on 200 m, May, 1954.

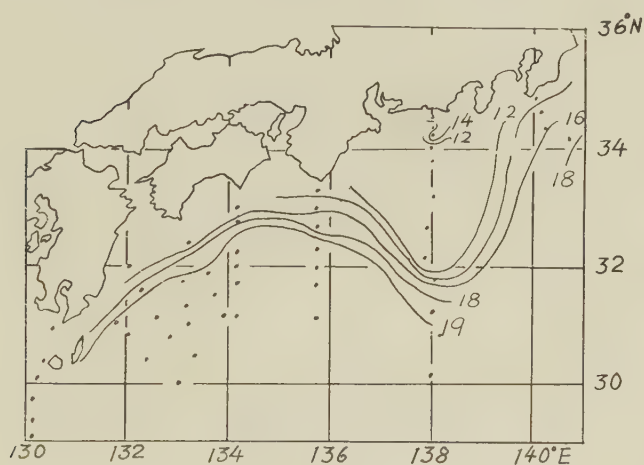


Fig. 11-10. Isotherms on 200 m, Nov. 1954.

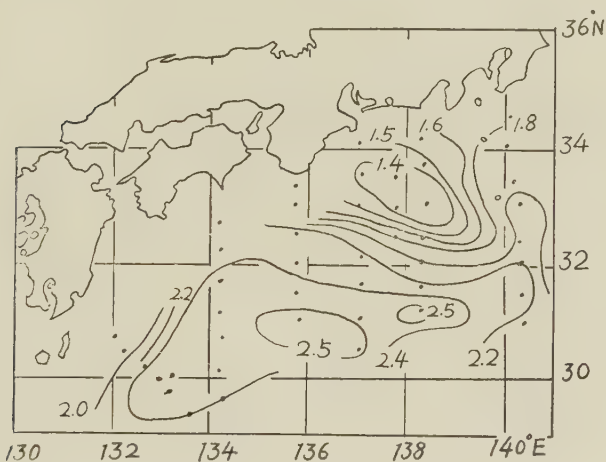


Fig. 11-11. Dyn. top., Aug. 1955. (0-1000db, anomaly)

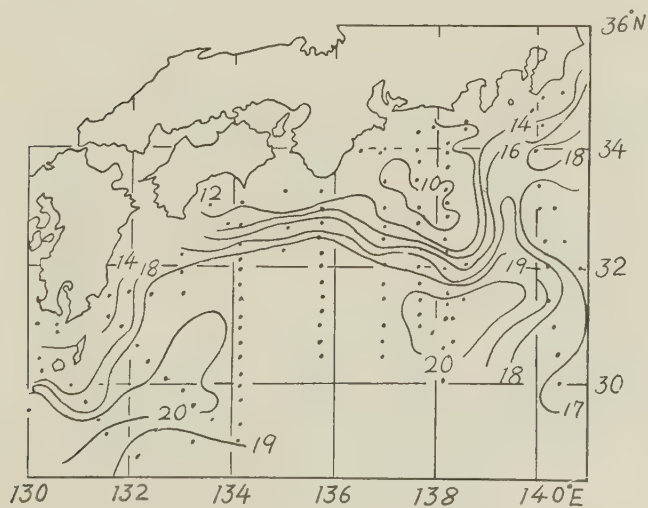


Fig. 11-12. Isotherms on 200 m, Aug., 1955.

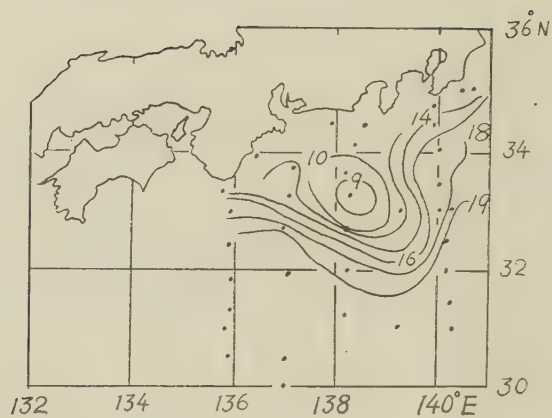


Fig. 11-13. Isotherms on 200 m, Oct.~Nov. 1955.

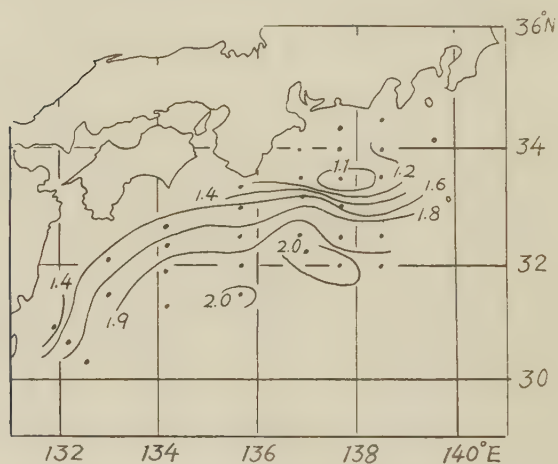


Fig. 11-14. Dyn. top., Mar. 1956. (0-800db, anomaly)

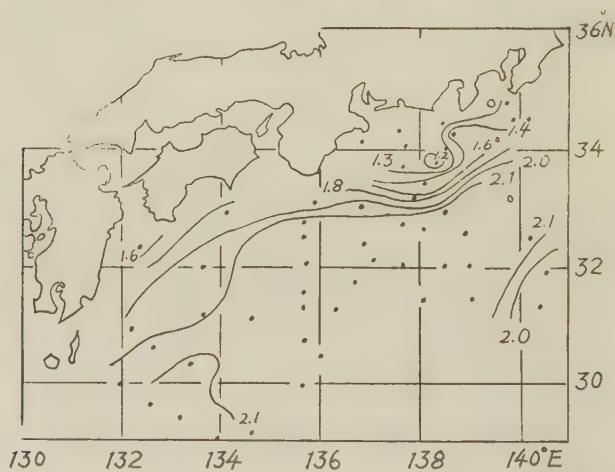


Fig. 11-15. Dyn. top. Aug. 1956. (0-800db, anomaly)

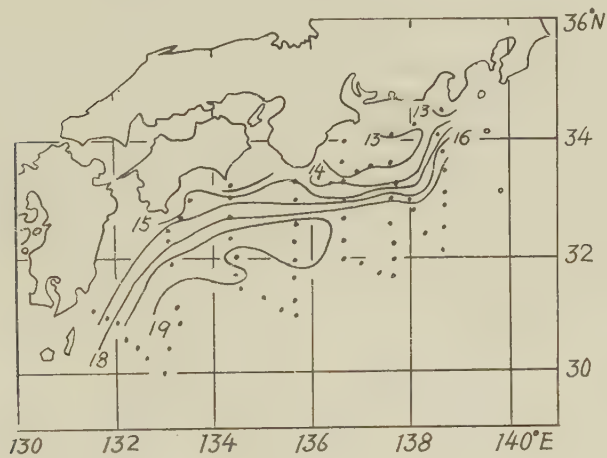


Fig. 11-16. Isotherms on 200 m, May 1957.

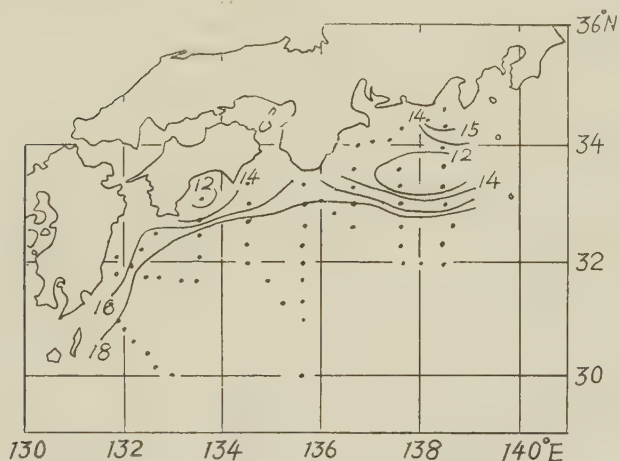


Fig. 11-17. Isotherms on 200 m, Nov.~Dec. 1957.

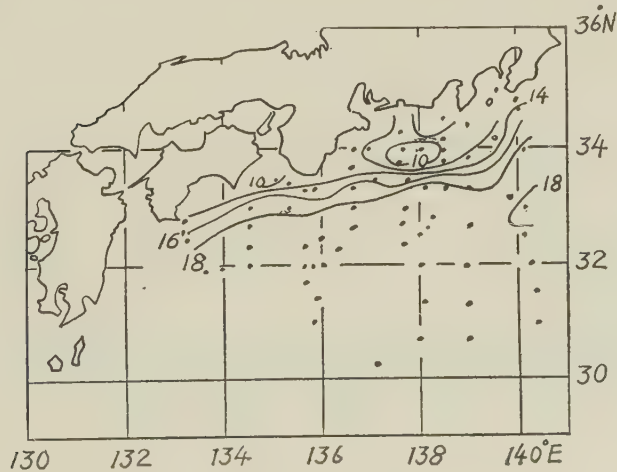


Fig. 11-18. Isotherms on 200 m, Mar. 1958.

Table 2. The condition of Kuroshio south of Ashizurizaki since 1952.

1952	June	Branch of the Kuroshio approaches Shikoku
	Sept.	Branch of the Kuroshio flows widely into the center of the Tosa Bay
	Oct.	Kuroshio approaches Ashizurizaki (17 nautical miles)
	Dec.	Kuroshio approaches Ashizurizaki but is not dominant.
1953	Feb.	Branch of the Kuroshio approaches near Shikoku
	Mar.	Velocity of the Kuroshio 3kts, 30~40 N. M. south of Ashizurizaki
	Apr.	Branch of the Kuroshio approaches extremily Shikoku
	July	Branch of the Kuroshio flows into the Tosa Bay
	Nov.	Kuroshio approaching is not seen
	Dec.	Kuroshio recedes offshore south of Ashizurizaki, and it does not approaches Shikoku in November and December.

formed to the west of Hachijo-jima. Besides, we can see the cold water mass in April 1948, similar to that in June 1947. In February¹⁰⁾ 1952 the cold water mass made an abrupt appearance in the Enshunada, and then in May the scale of this cold water mass was reduced. T. ICHIYE¹¹⁾ explained the appearance of this cold water mass as an effect of meteorology. So far as the author knows, we can not notice the abnormal states of the Kuroshio near Shikoku before the appearance of the cold water mass, and from a study of the life of the cold water masses of June 1947, of April 1948 and of February 1952, it seems that the characteristics of these water masses are different from those of the large persistent cold water mass stated above. T. ICHIYE¹²⁾ also stated that the Kuroshio axis south of Shionomisaki in July 1953 are offshore by utilizing the oceanographical observations along G line. It is very difficult to determine whether the offshoring of the Kuroshio axis is due to the existence of the large cold water mass in the offing of the Enshunada or to the abnormal mode of the Kuroshio south of Shikoku. If this offshoring of the Kuroshio axis relates to the receding of the Kuroshio south of Shikoku, this phenomenon is in accordance with the presage of the large cold water mass stated above.

In November—December 1958, the meandering of the Kuroshio¹³⁾ is very curious and small cold water mass are found to the east of the Kii peninsula and to the east of Miyakejima. The author¹⁴⁾ thinks that the cold water mass east of Miyakejima is made by the effect of the Izu submarine ridge. In view of such appearances of cold water masses in the sea area of the Kuroshio, we can conclude that the explanation of the mechanism of the cold water mass appearance is very difficult.

Although the difference between these cold water masses, one small and temporary, the other large and persistent, must be by all means studied, the problem is not easy and remains to be solved in future.

5. Speculation about the occurrence of the large cold water mass

When we study the dynamical topography (current chart) we can see in the large cold water mass a wavy pattern which has distinct wave lengths along the Kuroshio. These wave lengths are characteristic and their mean length is given in Fig. 12. The fact that this mean wave length may be stable is possibly supposed by the persistency of the above-stated phenomena.

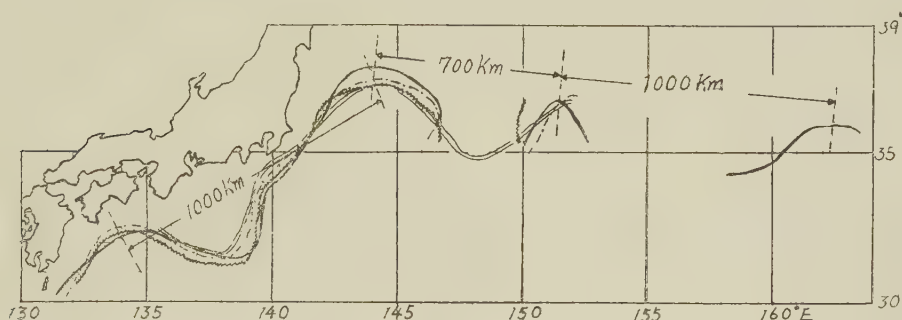


Fig. 12. Wave length of the Kuroshio flow pattern.
----- Aug. 1939, ... Aug. 1954, — Aug. 1955, — Aug. 1959

Why is this wave length stable? Sufficient explanation cannot be given to this phenomenon, but the author believes that this phenomenon is in outline explained as Rossby wave.

For standing wave pattern we think that progression speed of this Rossby wave¹⁵⁾ is zero.

$$L = 2\pi\sqrt{U/\beta} \quad L: \text{wave length}$$

Now we must study the mean velocity of the Kuroshio (U). The mean velocity of the Kuroshio in the range from south of Shikoku to east of the Boso peninsula may be considered as 30 cm/s–50 cm/s for the water column of 1000-m depth. The wave lengths corresponding to velocities of 30 cm/s–50 cm/s are about 800 and 1000 km, respectively ($\beta = 1.875 \times 10^{-13}$ (35°N)). These calculated wave lengths may be consistent with observations (Fig. 12). Though such a wave is stable, the conditions which lead to the stableness of the pattern must be analyzed. According to the analysis stated above, the first disturbance of the Kuroshio meandering may be recognized to the south of Shikoku, namely in 1933–1934 and in 1959. So we would like to analyze the disturbance of the Kuroshio south of Shikoku. As such problems are analyzed only by use of plentiful and systematic data, it is most suitable to use the data of oceanographical observations in 1959, because in 1934 and 1954 oceanographical observations were made less widely and less systematically than in 1959.

In January 1959 the Kuroshio transport was most abundant since 1954, according to the routine oceanographical observations by Nagasaki Marine Observatory in the East China Sea¹⁶⁾. The value is about 50×10^6 m³/s, one and half times as much as the mean value. In consideration of the width of the current, the velocity may be about 1.7 times as much as the mean. In case of the strong Kuroshio, the northeastward currents in the sea east of the Ryukyu Islands, which are always observed in normal years, are not noticed. As the current velocity below 500-m depth in this area is low, it can be assumed that the Kuroshio in the East China Sea reaches the 500-m depth. In the present paper also it is considered that the lower boundary of the Kuroshio is the 500-m depth.

When we study the mean current velocity of a water column of 500-m depth, the value is from 80 cm/s to 120 cm/s in the East China Sea. From the observations in January 1959, a current velocity of 120 cm/s is not unexpected. Now we take it into consideration that this Kuroshio traverses over the Ryukyu submarine ridge. The depth of the submarine ridge between Yakushima and Amamioshima is about 400 m. So the Kuroshio must traverse over a mound 100 m high.

When we study the case where a current traverses over a mountain, we may adopt the following formula in a simple form¹⁷⁾:

$$\sin \alpha = -\frac{f}{cD} \int_{x_1}^{x_2} h dx$$

$$\Delta x = x_2 - x_1$$

where c is the current velocity, f Coriolis parameter, h the mountain height, Δx the width of submarine ridge, D the current thickness in an undisturbed state, and α the deflecting angle when the current traverses over the top of the mountain (Table 3). Next, when we study the mean current velocity of a water column of 1000 m depth, the height of submarine ridge is 600 m, and the critical velocity of current (C_{cr}) over the submarine ridge is gained by following formula¹⁷⁾

Table 3. Deflecting angle (α) and critical current velocity (C_{cr}) over the top of submarine ridge.

water column of 500 m depth		water column of 1000 m depth
C	α	C_{cr}
120 cm/s	-19°	210 cm/s
80 cm/s	-34°	
$f=7.292 \times 10^{-5}$ (30°N)		$f=7.292 \times 10^{-5}$ (30°N)
$D=500$ m		$D=1000$ m
$h=100$ m		$h=600$ m
$\Delta x=30$ km		$\Delta x=100$ km

$$C_{cr} = \frac{f}{D} \int_{x_1}^{x_2} h dx$$

From this table it is known that the deflecting direction across the submarine ridge is remarkable when the ocean current velocity is weak. Such phenomena are often observed near Yakushima. In case of a strong current the direction is not so deflected. When we consider a water column of 1000-m depth, the critical velocity is very large and nonsense (Table 3). So it is thought that the remarkable effect of the ocean current may reach down to the 500-m depth. In meteorology when the airflow traverses over a mountain, the lee wave is produced. In oceanography, we must study the structure of what corresponds to this lee wave when the ocean current flows over a submarine ridge.

In the mean state of the Kuroshio in the East China Sea the northeastward current to the east of the Ryukyu Islands is fairly predominant, but in the case of strong Kuroshio the northeastward current is not developed (for example, in January 1959). In the former case the northeastward current approaches Shikoku in company with the meandering Kuroshio through the channel between Yakushima and Amamioshima. Meanwhile, when the Kuroshio is strong, there is no northeastward current to the east of the Ryukyu Islands, and the strong ocean current coming from the sea region between Yakushima and Amamioshima has a trend of straight flowing from west to east (Fig. 13), so that the Kuroshio flows almost eastnortheastward, and then the axis of the Kuroshio trends to recede offshore to the south of Shikoku.

The above analysis is only a speculation based on the oceanographical observations of 1959. Various analyses about the mechanism of the cold water mass have been offered by many oceanographers. T. NANNITI,¹⁸⁾ pointed out the development of the cold water mass southwestward from the sea region near the Izu Islands. The present author considers that at first the disturbance in the Kuroshio occurs south of Shikoku and moves eastward.

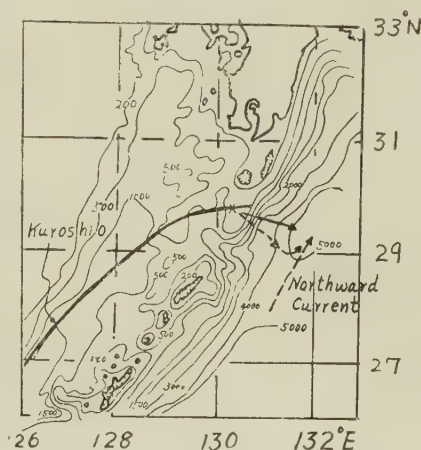


Fig. 13. Deflecting direction of ocean current over submarine ridge

[thin line:
bathymetric contour]
→ strong current
---→ weak current

6. Warm water mass in the Japan Sea

Hydrographical conditions in the Japan Sea have been also studied by many oceanographers in our country since thirty years ago. But when we now consider the distribution of water masses in the Japan Sea in detail, the studies heretofore are not sufficient. Because oceanographical observations in the past were not carried out at a sufficient large number of observing points, and the observations traversing the Tsushima current are scanty. In the present paper a rather different idea for the Tsushima current is expressed, an idea different from what was heretofore said usually. The present author shows chiefly the analysis of the data from 1950 to 1959, but for comparison's sake the data before the 2nd World War, the oceanographical observations in 1932, 1933 and 1941 are utilized.

Current charts (dynamical topography) and isothermal maps for the period from 1950 to 1959 are shown in Fig. 14. (these figures are often quoted in this paper.) It is said that the Tsushima current in the Japan Sea has three branches.¹⁹⁾ Namely, the first branch flows near Honshu in the sea area between Oki Island and Honshu, and the third branch streams most offshore. The second branch flows, making a detour of the Oki submarine ridge.

Though this idea may be generally right, the second branch north of the Oki submarine ridge shows preponderance, if we make reference to the data of 1950-1959. Now we can not discuss the third branch, because the oceano-

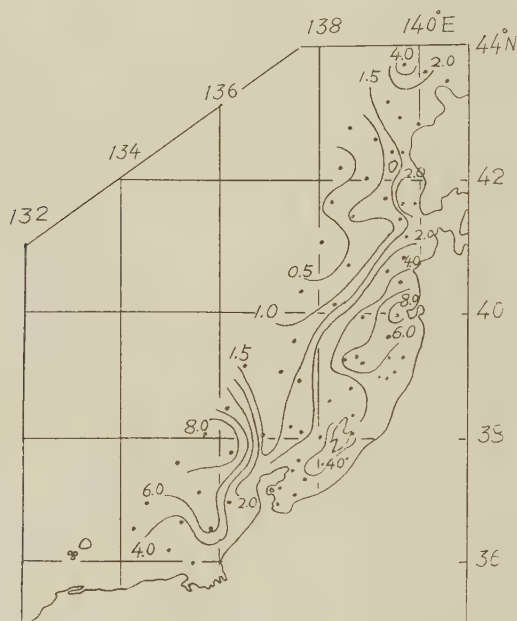


Fig. 14-1. Isotherms on 200 m Aug.~Sept. 1950.

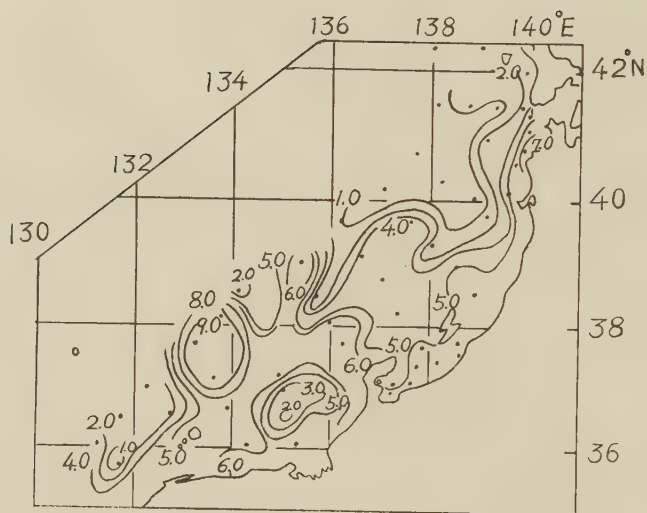


Fig. 14-4. Isotherms on 200 m, May~June 1952.

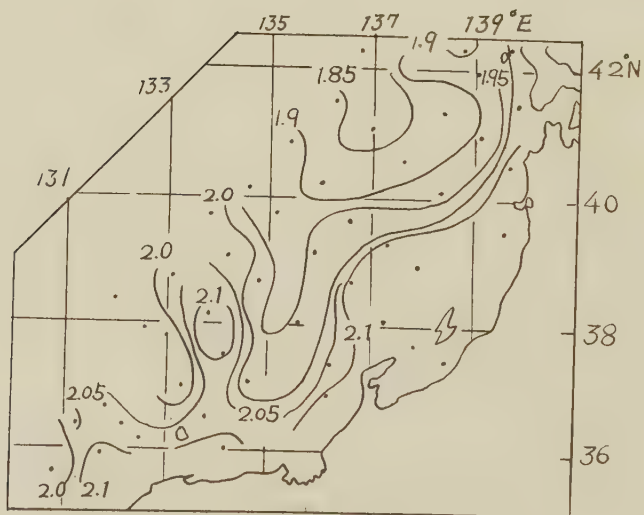


Fig. 14-5. Dyn. top. May~July, 1953.
(0-400db) (Hydro Dep.)

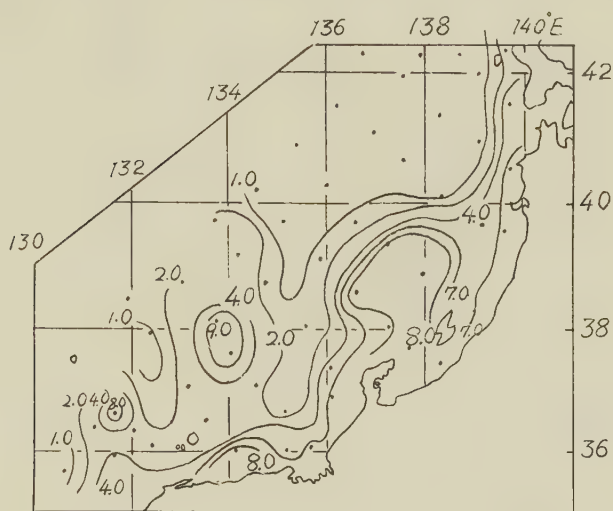


Fig. 14-6. Isotherms on 200 m, May~July, 1953.

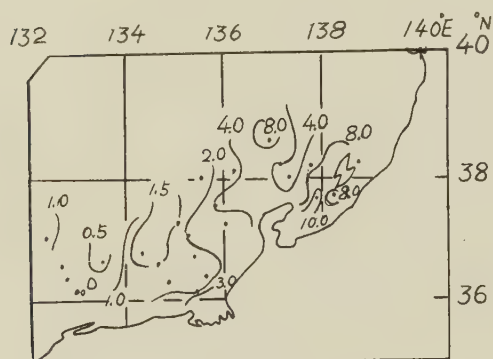


Fig. 14-7. Isotherms on 200 m, Nov. 1953.

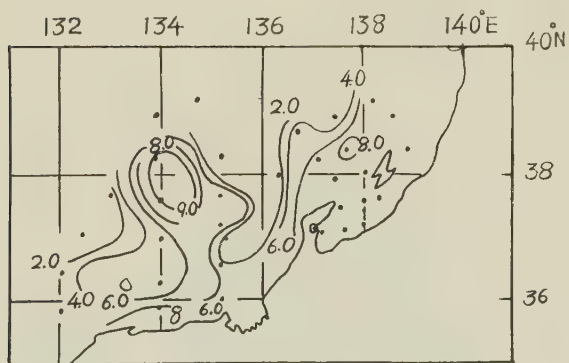


Fig. 14-8. Isotherms on 200 m, Mar., 1955.

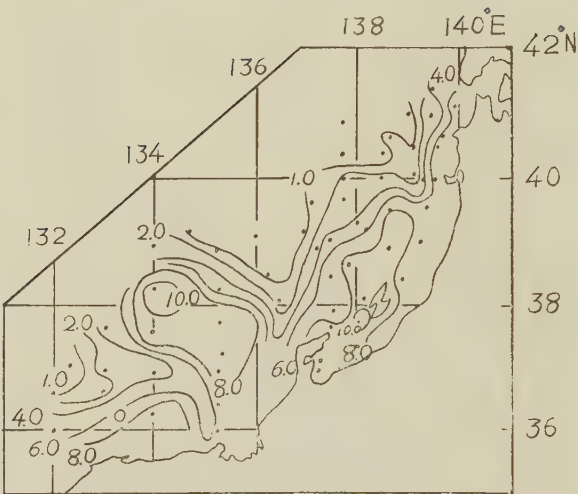


Fig. 14-9. Isotherms on 200 m, July 1955.

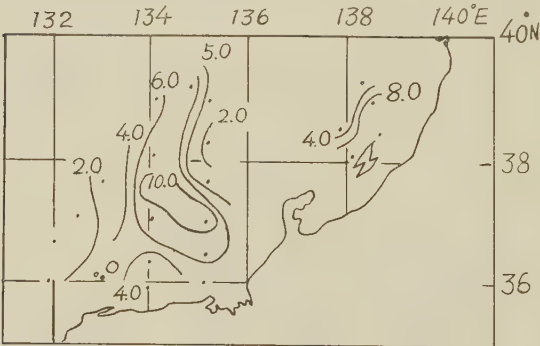


Fig. 14 10. Isotherms on 200 m, Feb., 1956.

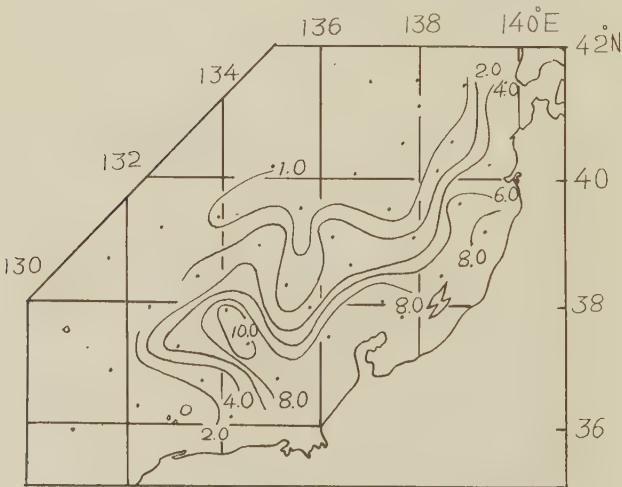


Fig. 14-11. Isotherms on 200 m, May 1956.

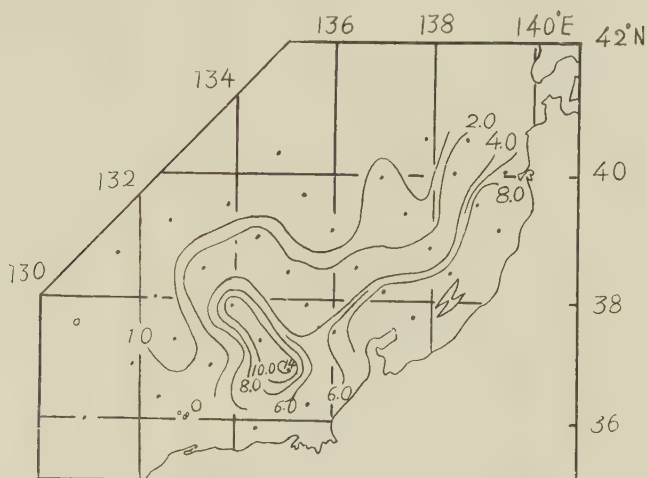


Fig. 14-12. Isotherms on 200 m, Aug., 1956.

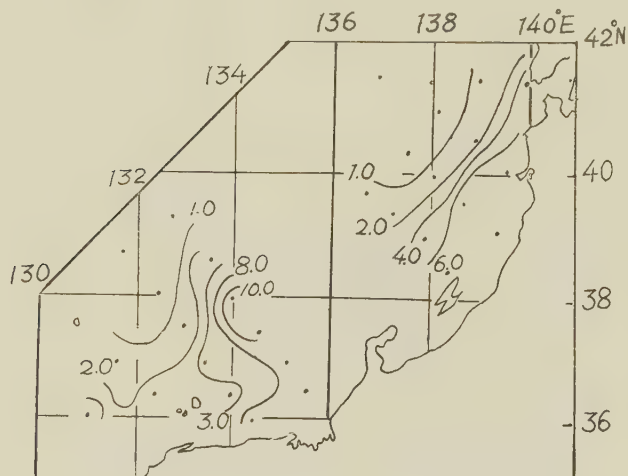


Fig. 14-13. Isotherms on 200 m, Oct~Nov., 1956.

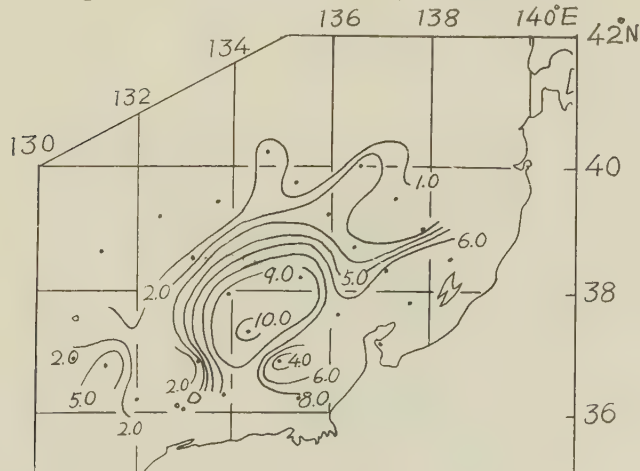


Fig. 14-14. Isotherms on 200 m, Mar., 1957.

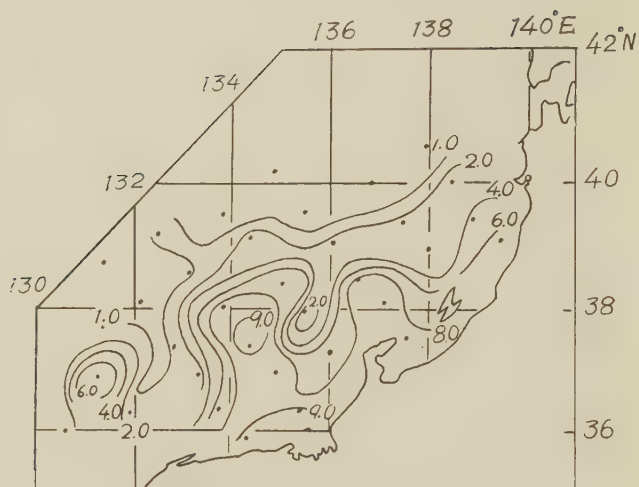


Fig. 14-15. Isotherms on 200 m, May, 1957.

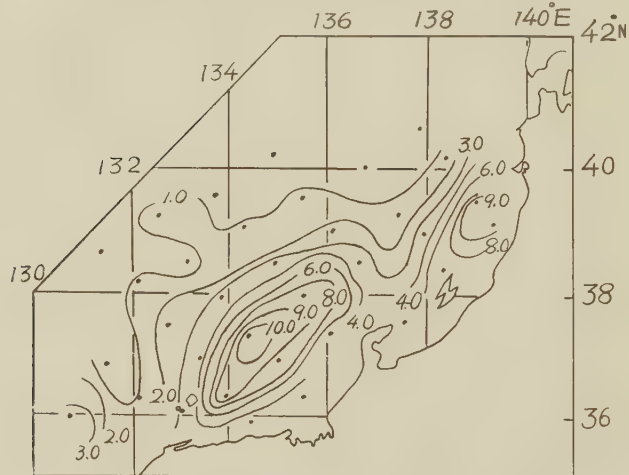


Fig. 14-16. Isotherms on 200 m, Aug., 1957.

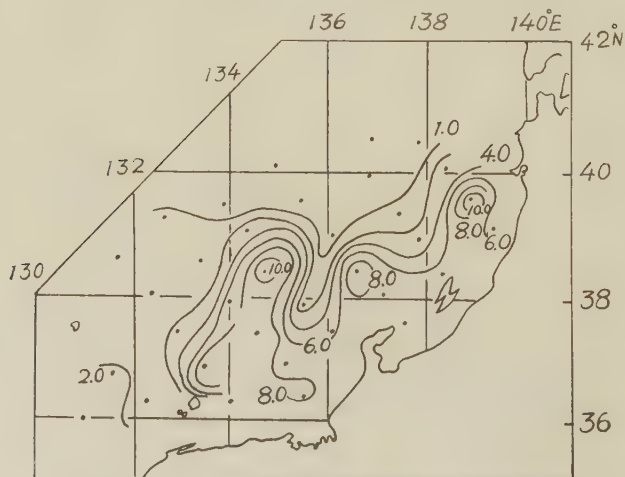


Fig. 14-17. Isotherms on 200 m, Nov., 1957.

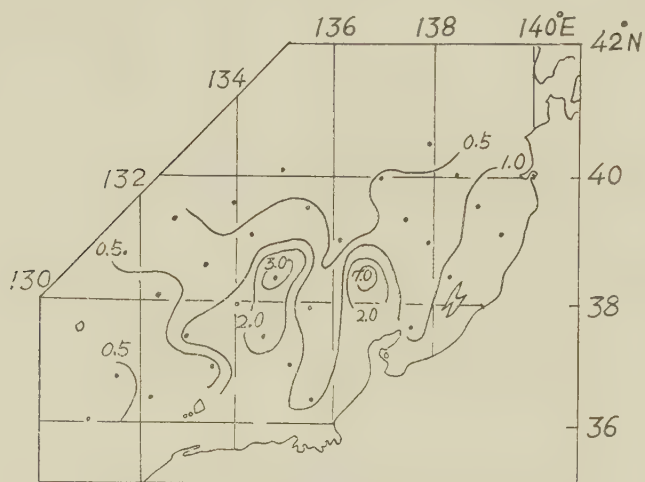


Fig. 14-18. Isotherms on 300 m, Nov., 1957.

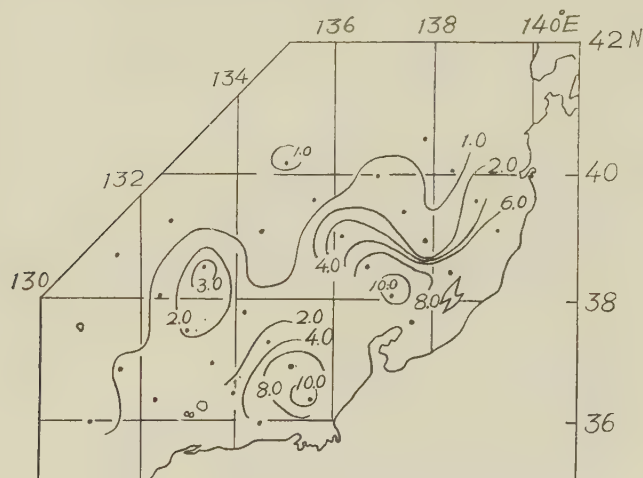


Fig. 14-19. Isotherms on 200 m, Mar., 1958.

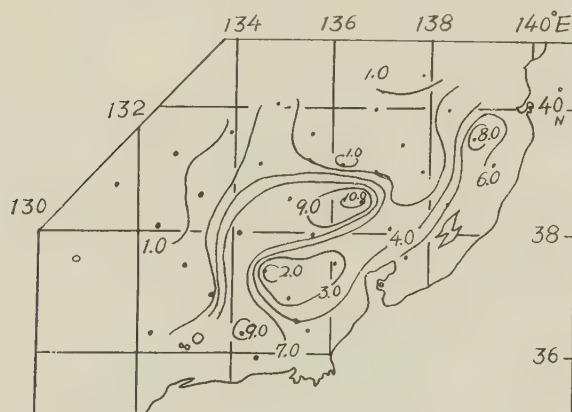


Fig. 14-20. Isotherms on 200 m, May, 1958.

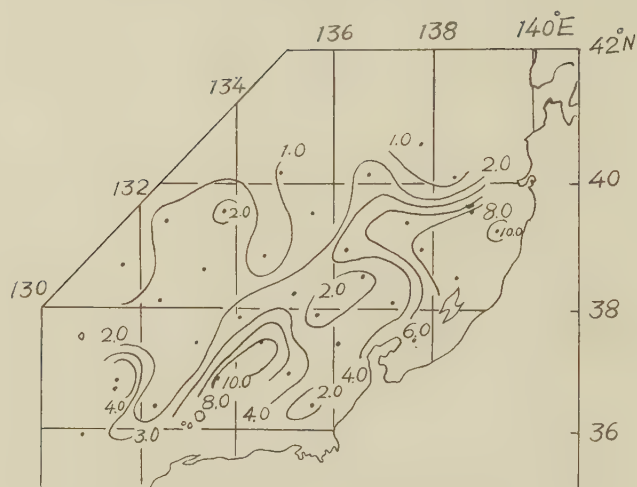


Fig. 14-21. Isotherms on 200 m, Aug., 1958.

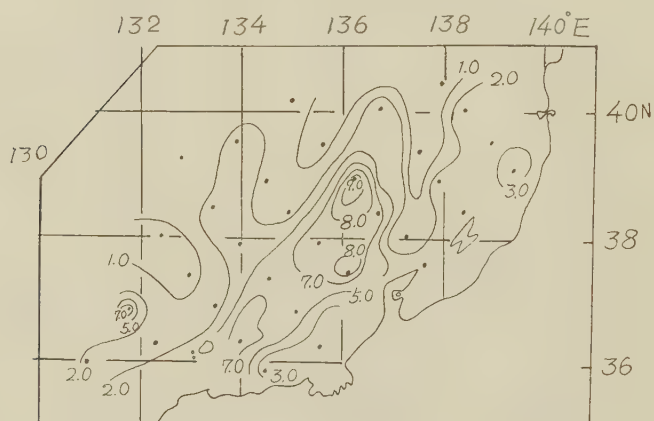


Fig. 14-22. Isotherms on 200 m, Mar. 1959.

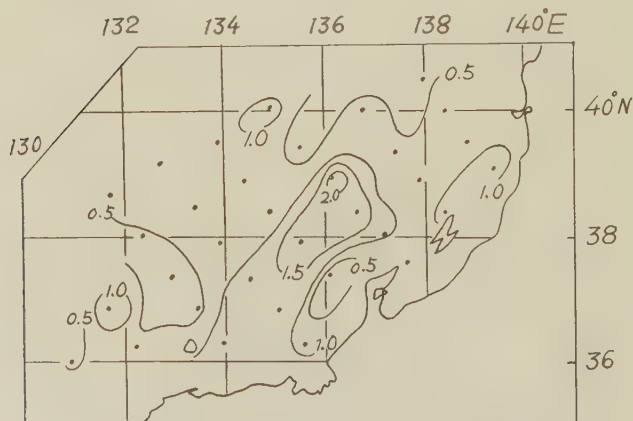


Fig. 14 23. Isotherms on 300 m, Mar., 1959.

graphical observations are not carried out in the sea area of this branch for political reasons.

When we compare the first branch with the second branch for the data from 1955 to 1959 (refer to Fig. 14), it is reasonable that the northward component north of Oki Island is most dominant.²⁰⁾ Next, in order to simplify the current chart of dynamical topography, adoption of the isothermal map at 200-m layer as a substitute of the chart of dynamical topography is useful and convenient (the similarity between the dynamical topography and the isothermal map at 200-m layer is shown in Fig. 14). So we use the isothermal maps at 200-m layer as a method of analysis of current and water mass.

According to the researches hitherto, the vertical distribution of the Tsushima current shows three types of water, namely, waters of surface layer, intermediate layer and deep layer. The water at 300-m layer in the Japan Sea is generally said to have characteristics of a deep water, but according to this study, intermediate water occupies even the 300-m layer in the warm water mass area. Of course the absolute values of water temperature at 300-m layer are changed yearly. But the water of relatively high temperature is in existence in the warm water mass, which is indicated by the isotherms at 200-m layer.

It is reasonable that the area of high water temperature is produced by sinking motion, but the yearly difference of water temperature at 300-m layer may be due to the yearly difference of sinking motion. In Fig. 15 the areas of warm water mass are shown by a close isotherm of 9°C north of Oki Island, by a closed isotherm of 4°C north of Hamada and by a closed isotherm

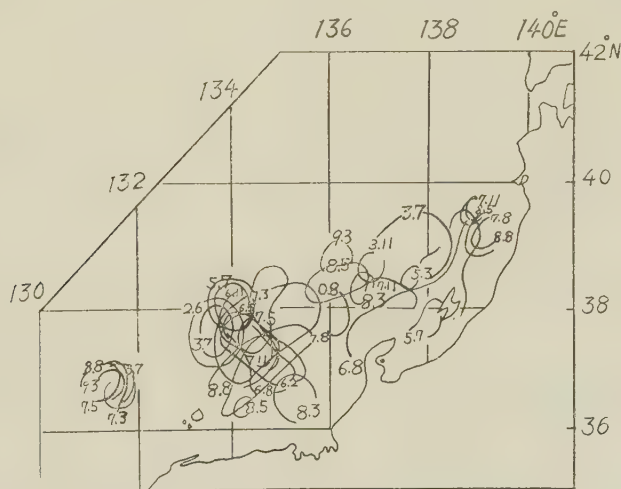


Fig. 15 Locations of warm water mass.
0.8: observed Aug. 1950
7.3: „ Mar. 1957

of 8°C north of the Noto peninsula. From this figure the warm water mass north of Oki Island is always in existence and its location is almost constant.

The warm water mass north of Hamada is not always seen from 1950 to 1959, but if it once appears, its location is constant.

On the other hand the warm water mass north of the Noto peninsula changes its location rather yearly, but to the west of Akita the location of warm water mass is constant.

From these distributions of warm water mass, especially from the warm water mass north of Oki Island, we can estimate a close relation between warm water mass and bottom topography. And an investigation on the bottom topography of the Japan Sea gives Fig. 16. Comparing Fig. 15 with Fig.

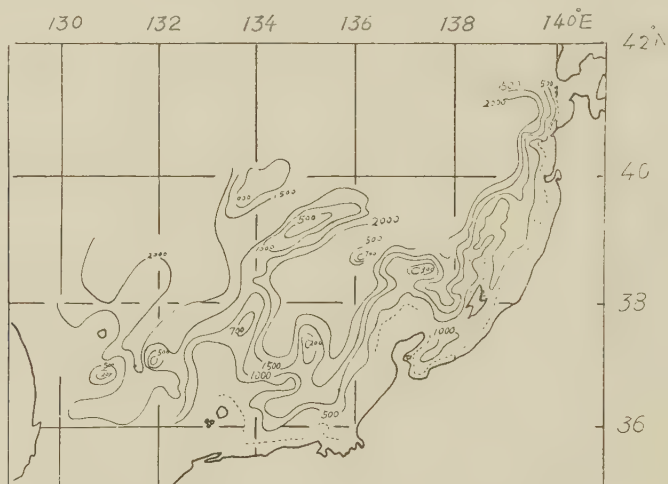


Fig. 16. Bathymetric chart in the Japan Sea.

16, we can see that the warm water mass occupies the area behind the submarine ridge and sea mount. From 1953 to 1958 the warm water mass can be traced down to the 400-m layer. From these facts we can estimate the sinking motion will reach the 400-m depth.

The warm water mass north of Hamada (near Takeshima) will be also related with the bottom topography. Even if we utilize the sounding data on the charts which are now published, the isobath contour can not be precisely drawn because of the scantiness of sounding data. In order to study the relation between the distribution of water mass and the bottom topography, it is necessary to make an accurate bathymetric chart. So the author hopes that sounding would be frequently carried out in the Japan Sea.

Next, we will study the warm water mass north of the Noto peninsula. The location of this water mass seems to be change more often than the locations of the both warm water masses stated above. But the warm water

masses in March 1959 and in May 1958 may be connected with the effect of the Yamato Bank and its related bank. In short, by comparing the distribution of water mass with the bottom topography, we can make an analysis of hydrographical conditions in the Japan Sea, and further, we shall be able to conclude that the warm water mass is controlled by the bottom topography.

It is noticeable, however, that even the most stable warm water mass north of Oki Island changes its location in special years. In August 1950 the warm water mass north of Oki Island shifts more eastward than in a normal year. In March 1959 the same warm water mass moves more northeastward than in a normal year.

It is necessary to study why the location of warm water mass indicates such differences from special years. In Fig. 16 we can see that the Tsushima current reaches the Yamato Bank along the Oki submarine ridge, changes its direction on the west side of the bank, traverses over this ridge and the related sea mount, and then the warm water mass north of the Noto peninsula appears. It is seen that the characteristic of the Tsushima current in March 1959 is its extreme northing.

In August 1950 the warm water mass north of Oki Island is located more easterly than in normal year. The cause of this cannot be clarified, for we have no data of oceanographical observations in the west of the Japan Sea for that year, and so the accurate conditions of the Tsushima current are unknown to us.

In November 1953 the warm water mass is seen only to the north of the Noto peninsula, and the sea area north of Oki Island is occupied by the water of low temperatures at the 200-m layer, utterly different from normal year conditions. (This hydrographical condition is similar to that of May—June, 1941, and this problem will be stated later.)

In 1958 the warm water mass north of Oki Island is location slightly nearer to land than its normal state of 1955 to 1957. Next, we would like to study the warm water mass north of Hamada. This warm water mass is not always in existence, but once it appears, it occupies the constant space. Because of the scantiness of data a reasonable conclusion can not be drawn, but the current chart by GEK indicates an anticyclonic circulation in the warm water mass. Making reference to Figs. 16 and 14-6, 14-14, this anticyclonic circulation may be said to correspond to the series of sea mounts near Ullung Island and Takeshima.

The author has deduced that the appearance of the anticyclonic circulation (in other words, the appearance of warm water mass) might have a close connection to the direction and velocity of the Tsushima current near the Tsushima Straits. But the data for elaboration of this deduction are few. In the near future it will be necessary to carry out continuous oceanographical observations near the Tsushima Straits and to find some relation between the

variations of the Tsushima current and the appearance of the warm water mass near Takeshima.

In the above, only the warm water mass is studied. But the research about the distribution of water mass should correspond to the study of meandering of ocean current. For the Tsushima current, we can determine its main wave length to be 300 km. This wave length also may be controlled by the bottom topography.

In the above the hydrographical conditions have been discussed from a view point of the distribution of warm water mass. Here we would like to study the hydrographical conditions before the 2nd World War. Isothermal maps of May—June 1941 are shown in Fig. 17. Taking a glance at this

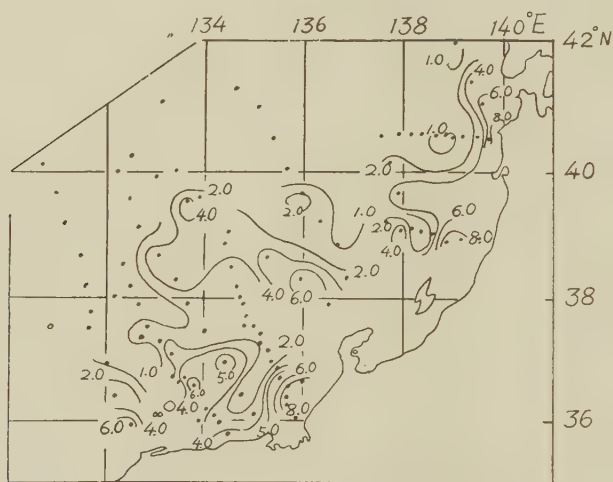


Fig. 17. Isotherms on 200 m, May~June 1941.

map, we can see the difference of hydrographical conditions between the year 1941 and the years 1955 to 1957. Namely, in 1941 the warm water mass near Oki Island is small and the sea area of Oki Island is occupied by the water of low temperatures at the 200-m layer. As the low water temperature area is wide, viewing from the distribution of water temperature at the 200-m and 300-m layers (Fig. 17), the sinking motion does not reach so deep, and it is seen that the ocean current has an effect upon rather shallow layers. It is supposed that the first branch of the Tsushima current is dominant and the second branch is not developed.

From the oceanographical observations of 1929²¹⁾ and 1930²²⁾, the warm water mass near Ullung Island exists distinctly, but the warm water mass north of Oki Island is not found. Owing to the scantiness of oceanographical data, it can not be affirmed that the warm water mass north of Oki Island does not occupy the space entirely. Investigating the oceanographical obser-

uations in June 1932 and August of 1933, we can notice that the warm water mass north of Oki Island in these years is very similar to that in 1955—1957.

It is generally said that the hydrographical conditions in the Japan Sea before the 2nd World War are quite different from those after the War. From a view point of the distribution of water mass, however, the hydrographical conditions of 1932 and 1933 could be said to be similar to those of 1955 to 1957.

7. Future problem for clarifying the occurrence of the cold and warm water masses, and conclusion

One of the easiest and most effective methods which we can adopt for clarifying the mechanism of the large cold water mass in the Pacific Ocean and the warm water mass in the Japan Sea may be collect and arrange the reports on ocean currents from merchant ships and fishing boats. Unfortunately we have not the reports on ocean currents at all, because no institution at present is not gathering the reports on ocean current. If we collect and arrange such reports, they should offer us continuous information about abnormal states of the currents. For example, if the hydrographical condition of the Tsusoima current is very anomalous May—June 1941, the anomaly should be more distinctly known by the reports on ocean current. So the author strongly hopes that the reports on ocean currents should be collected and then up-to-date charts of ocean currents should be prepared. And it is an important problem in future to know how the effects of submarine ridges will be manifested in various flow patterns of the Kuroshio and the Tsushima current, by using an electronic computer. Such systems are one of the best methods to clarify the mechanism of the cold and warm water masses.

Acknowledgements

It is a large pleasure to the author to express his hearty thanks to Dr. Y. TAKENOUTI, Dr. T. NANNITI, Dr. Y. SAITO, Dr. K. HISHIDA and Dr. K. YOSHIDA for their encouragements and advices, to Mr. K. WADA for his assistance in the polishment of this paper, and to Miss K. OKADA for her toil of drawing the figures of this paper. Besides, the author wishes to express his many thanks to observers of research ships of Japan Meteorological Agency, Hydrographic Department and Fisheries Agency.

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Some Calculations on the Equatorial Circulation

By

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1. Introduction

It seems that there may still be various different theoretical models which could account for the observed features of the Equatorial Undercurrent (Cromwell Current) in the eastern Pacific. Physical significance of the models and hence the validity of the theories may more critically be tested by comparing those results with observations of the current structures under different conditions. As J.A. KNAUSS suggested (personal letters, 1960), the SCOR-sponsored international oceanographic survey of the Indian Ocean may be expected to provide useful information of the current and will help somehow to test the theories proposed for the Pacific Undercurrent. If the easterly wind stress over the equatorial region is responsible for the Cromwell Current as described by the present author (YOSHIDA, 1959*), there would be no eastward undercurrent in the Indian Ocean where the equatorial wind is from the west during most part of the year. On the other hand, if, as also suggested in [Y], the undercurrent is a phenomenon characteristic to the equator with a singularity of vanishing vorticity due to the Earth's rotation, the eastward winds along the equator of the Indian Ocean may produce different type of undercurrent flowing westward in opposite direction to the surface eastward current. The winds over the equatorial regions in the Indian Ocean vary from latitude to latitude more widely than in other oceans, and they also reveal marked seasonal variation. With a modification by taking account of the latitudinal variation in the winds, the same theoretical model as employed for the Pacific is attempted to apply to the Indian Ocean.

Another problem considered in this paper is concerned with a South Equatorial Countercurrent in the Pacific of which an evidence has recently been shown by J. REID. With this newly discovered current, though not yet fully confirmed by direct measurements, the equatorial current system in the Pacific seems to have increased the degree of its symmetry about the equator. Such a symmetric structure of the equatorial currents may be anticipated from an approximate symmetry in the distributions of the zonal wind stress about the equator. A rough calculation from wind stress data may suggested a possibi-

* This paper will be referred to as [Y] in the present article.

lity of a real existence of this current.

2. The Equatorial Undercurrent in the Indian Ocean

The wind conditions over the equatorial areas of the Indian Ocean differ entirely from those in the Pacific and Atlantic. The wind stress data (HIDAKA, 1958) indicate that the annual mean winds at the equator are eastward at nearly all longitudes in the Indian Ocean. Although the winds at the equator reveal considerable seasonal variations, they still remain eastward during all the seasons except for western edge portions and for western part during winter. Furthermore, the variations in the wind stress with latitudes are quite appreciable and may not necessarily be negligible for the Indian Ocean.

If the basic ideas governing the author's theoretical derivation for the Cromwell Current in the Pacific [Y] should remain valid for the Indian Ocean with different winds, certain conclusions could be obtained as given in the following.

The equations (3.1) to (3.11) and (4.1) to (4.7) in [Y] do not depend upon any functional form of $\tau_x(y)$. That is, these equations will remain valid even though τ_x varies largely with y , where τ_x is the eastwest component of the surface wind stress and y is the coordinate taken positive northward from the equator.

The maximum east-west velocity at the very equator in the layer just below the surface layer, denoted by the same symbol, $u_M(0)$, can be known if the value of the zonal wind stress at the equator, $\tau_x(0)$, is known. From the first line of (4.8), (3.9) and (4.6), we have

$$u_M(0) = -\frac{6H}{5\pi A} \tau_x(0), \quad (1)$$

where H is the mean thickness of the surface layer and A is the vertical mixing coefficient.

From the data of mean zonal wind stress (HIDAKA, 1958) the annual mean wind at the equator of the Indian Ocean is from the west, 0.1 to 0.3 dynes cm^{-2} approximately. Then $\tau_x(0)$ is positive and consequently $u_M(0)$ must be negative from (1), indicating the eastward undercurrent. With the numerical values, $H=10^4$ cm and $A=20 \text{ cm}^2 \text{ sec}^{-1}$ as used in [Y], and $\tau_x=0.2$ dynes cm^{-2} , the maximum velocity becomes $u_M(0)=-40 \text{ cm sec}^{-1}$ corresponding to the westward current with the core at 200 meters. At the surface, the current directs eastward due to the eastward winds as well as the meridional pressure distribution.

In the present case for the Indian Ocean, the meridional variation in the zonal wind stress might have to be considered. In order to see if this variation might materially affect the meridional distributions of the velocities, we must solve (3.10) for certain realistic form of $\tau_x(y)$. To a first approximation,

it will be assumed that the meridional distribution of the wind stress near the equator may be expressed by

$$\tau_x(y) \cong \tau_0(\alpha_0 + \alpha_1 y + \alpha_2 y^2) \quad (2)$$

The equation (3.10) is valid but in this case τ_x is as given by (2). The boundary condition is that the first term of the left hand side of (3.10) tends to zero as we go sufficiently away from the equator. It is seen from the solution obtained that the latitudinal distributions of vertical velocity, $w_H(y)$, may depend only on the value of the wind stress at the equator and that of the curvature of the stress at the equator. If for instance, $\tau_x(y)$ varies linearly with latitudes with $\tau_x(0) = \tau$, the distribution of w_H will be the same to the case when τ_x is uniform everywhere with $\tau_x \equiv \tau$. The curvature of the wind stress at the equator will, on the other hand, affect the vertical velocity, but this effect is only to add constant negative or positive values to w_H for all latitudes, according to negative or positive values of $\partial^2 \tau_x / \partial y^2$ at the equator, respectively. The distribution of w_H is symmetrical about the equator, as far as the wind stress distributions can be expressed approximately by a quadratic function as assumed.

The numerical values of $\tau_x(0)$ and $\frac{\partial^2 \tau_x}{\partial y^2}$ for the equatorial regions in the Indian Ocean are computed from the wind stress data given by HIDAOKA. These rough calculations indicate that the effect of the curvature is almost negligible. At latitudes lower than about 2° N. and S., w_H is positive (sinking), and is negative (upwelling) higher than about these latitudes.

In contrast, there is an evidence for the equatorial upwelling from a Japanese observations in the eastern portion of the equatorial Indian Ocean which were taken during January. Curiously, the vertical section prepared by YAMANAKA (YAMANAKA, 1958) seems to indicate clearly the similar thermal feature across the equator as seen in the eastern Pacific. Further, even the equatorial undercurrent going eastward might be suspected from this temperature distribution revealed. The statistical wind data do not seem to give any account for this observed upwelling as far as the present theoretical model is applied. Unfortunately, wind conditions during the observations are not given in YAMANAKA's paper. Surface current charts (SVERDRUP et al, 1942) indicate the westward surface current at the equator during February to March. If this is true in actuality during these months, there may possibly be an eastward undercurrent as well as equatorial upwelling, perhaps exclusively during those months as indicated by YAMANAKA. A complete interpretation is then difficult as yet, if the winds are as given in HIDAOKA's statistical data.

3. A South Equatorial Countercurrent in the Pacific

Joseph L. REID, Jr. has recently shown an evidence of a South Equatorial Countercurrent in the Pacific Ocean, based on the calculations of geostrophic flow for cross-sections made by various past expeditions (REID, 1959). An eastward-flowing current as indicated by REID can be expected dynamically

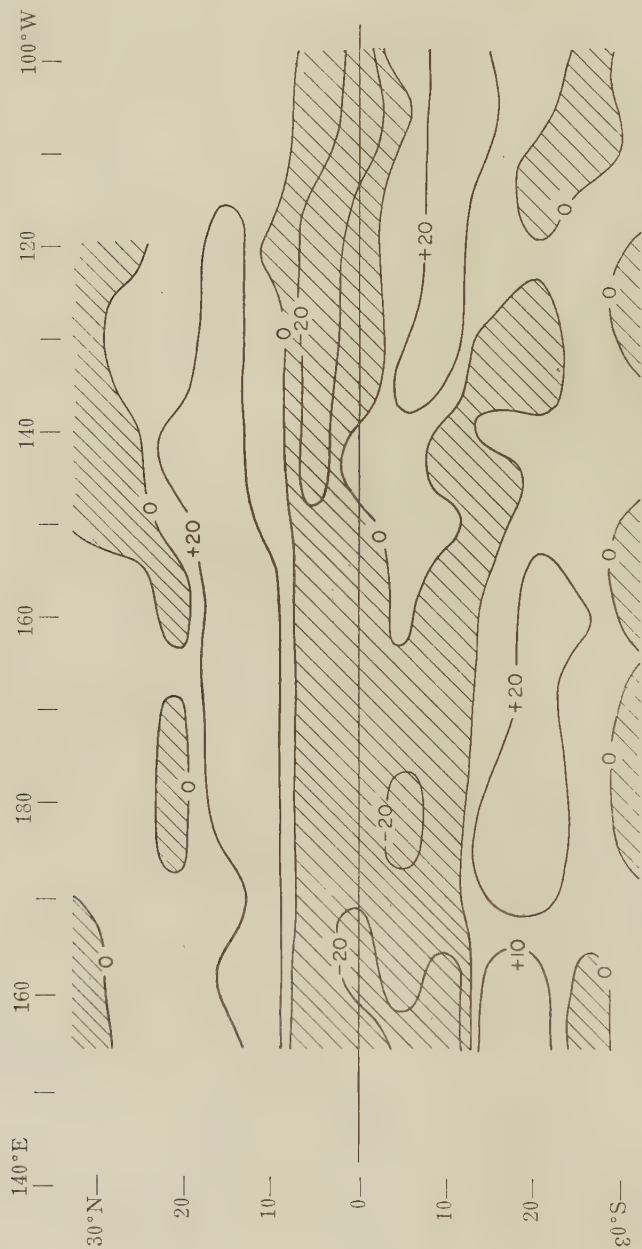


Fig. 1. The distribution of $\partial^2 \tau_x / \partial y^2$ (the annual mean values in the Pacific Ocean):
Unit for the numerical values is 4×10^{-18} c. g. s.

from the calculations of wind-driven transport. The wind stress data, on which the calculations are based, are again from HIDAOKA's paper (1958). This counter-current flows in the direction opposite to the local winds, easterly winds. Very roughly speaking, the regions of positive wind stress curl correspond to those of surface divergence with sea level low and thermocline shallow. Those of negative curl does to the surface convergence with sea level high and thermocline deep. Then in the region of negative values of $\partial^2\tau_x/\partial y^2$, the surface geostrophic flow will be directed eastward in the Southern Hemisphere. Somewhat smoothed features of the distributions of $\partial^2\tau_x/\partial y^2$ are illustrated in Fig. 1. The shaded areas indicating negative values of $\partial^2\tau_x/\partial y^2$ should correspond to the areas of surface eastward transport. Remarkable would be the regions of the predicted eastward transport extending down to 10 to 15 degrees south of the equator to the west of 140°W, and somewhat further south to the east. In the eastern portion, however, the above simple theory may not necessarily be valid. The negative values indicated along the equator will not be closely related to the actual surface current at the equator as the dynamical relationship given above will not be valid at and very near the equator.

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Determination of Uranium in Sea-Water Using a Polarographic Method

By

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Abstract

Polarographic method was applied for the determination of uranium in sea-water in the Pacific Ocean and the results ranging from 1.5 to 1.7 $\mu\text{g}/\text{L}$ of sea-water were obtained.

Introduction

Many investigations¹⁾ dealing with the determination of uranium in sea-water have been reported: most of them utilized fluorometric method of analysis and the results are ranging from 0.4 to 2.7 μg uranium per 1 L of sea-water. Activation analysis for the same purpose was applied recently by STEWART and others.²⁾ Polarographic method was attempted by HECHT et al.;³⁾ after concentrating uranium by means of anion-exchange resin, they determined uranium in artificial sea-water by the catalytic wave of uranyl ion in nitric acid medium, however, no data was shown for the natural sea-water.

ISHIBASHI and his co-workers,⁴⁾ recently, developed a polarographic method for the quantitative determination of small amount of uranium using Mordant Blue 2R, a di-o-hydroxyazo dye, as the reagent. In the present paper, this method was applied for the determination of uranium in sea-water collected in the Pacific Ocean off the coast of Shirahama in Wakayama Prefecture. The result obtained was 1.5-1.7 $\mu\text{g}/\text{L}$.

Experimental

Apparatus and reagents

Apparatus: Yanagimoto Pen Recording Polarograph, Model PA-102 was used. PH adjustment was made by the Toyo Denshi Kiki Co., pH meter.

Reagents: The solutions of MB2R, uranyl acetate, and Triton X-100 were prepared according to the procedures reported in the previous paper.⁴⁾ All the reagents used were analytical grade and the solvents used were redistilled when necessary.

Cellulose column: The column used was a burette of about 2 cm in diameter and 30 cm in length. The cellulose pulp (Toyo Roshi Co. filter wad) for the column was prepared by boiling it with 5% nitric acid for an hour to remove impurities and the pulp was then filtered and washed free from nitric acid with water. It is further washed with ethyl alcohol and finally with ether. After draining at a filter pump, the pulp is stored in a closed container and is ready for use.

Sample sea-water: The sample sea-water was collected in the ocean off the coast of Shirahama, Wakayama Prefecture, Japan on 20th of February, 1960.

Chlorinity: 19.09‰, pH=8.10.

Artificial sea-water: The artificial sea-water used was prepared according to the prescription of Lyman and Fleming.⁵⁾

Polarographic determination of uranium⁴⁾

Fig. 1, curve A shows the wave of 0.024 mM MB2R in 0.12 M sodium acetate, 0.001% Triton X-100 solution (adjusted to pH 5.3 ± 0.2 by the addition of perchloric acid). The azo functional group of MB2R is reduced at about -0.3 V (vs. SCE). Curve B in Fig. 1 is a polarogram of the same solution as that giving curve A, except that the solution contained $4 \mu\text{g/ml}$ of uranyl ion. The second wave appears at a potential about 0.15V more negative than that of the first wave. The total wave height increases with increasing uranyl concentration because the reduction of uranyl occurs together with that of the uranyl-MB2R complex. In spite of the simultaneous reduction of uranyl-MB2R and of the free uranyl ion, the height of the second wave is proportional to the uranyl concentration up to about $6 \mu\text{g/ml}$. The coexistence of

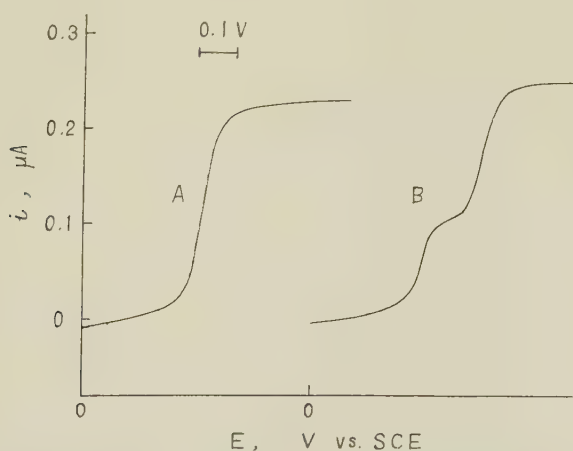


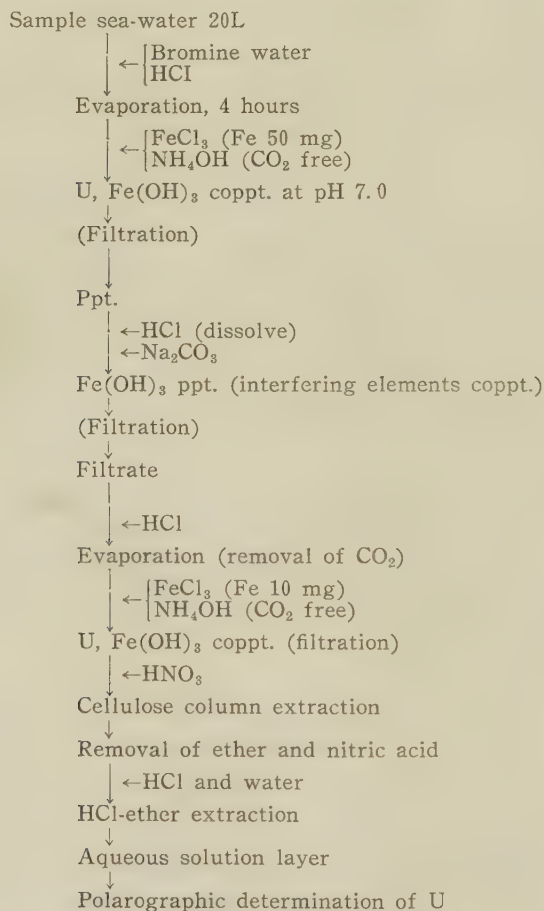
Fig. 1. Polarogram of MB2R alone (curve A) and mixture of MB2R and uranyl ion (curve B). See text.

Determination of uranium in sea-water

The determination of uranium dissolving in sea-water was carried out applying the polarographic method mentioned above after concentrating uranium from sea-water and separating the interfering elements. The scheme of the procedures are shown in Fig. 2.

Procedure A: To 20 litres of sea-water, is added 5 ml bromine water and 20 ml of concentrated hydrochloric acid. Then the solution is boiled in a large evaporation dish for about 4 hours until its volume decreases to about 15 liters in order to decompose organic substances and to remove carbon dioxide. The solution is then transferred to a bottle (20 liters in capacity) and ferric

Fig. 2. Schematic Diagram of the Method for the Determination of Uranium in Sea-water.



Final volume of the solution: 10.0ml
(0.024 mM MB2R, 0.12 M NaOAc, pH=5.3)

iron (50 mg Fe) is added. By the addition of carbonate free ammonia until the pH of the solution becomes 7.0 ± 0.2 , uranyl ion is coprecipitated with ferric hydroxide. After standing for two days, the precipitate of ferric hydroxide, Al, Fe, Co, Ni, Th, Ti, V, Pb, Cu, F, and PO_4^{3-} interferes the determination of uranium and the preliminary removal of them is necessary.

xide is collected by filtration and is washed with water to remove sodium chloride. In these processes, carbon dioxide is kept shut off as completely as possible. The precipitate is then dissolved with hydrochloric acid.

Procedure B: To this solution sodium carbonate (2 g in excess after ferric hydroxide is completely precipitated) is added and interfering elements are coprecipitated with ferric hydroxide. In this process uranyl ion does not coprecipitate with iron by the formation of soluble carbonate complex. The precipitate is filtered off and the carbonate is completely removed from the filtrate by the addition of HCl. The dissolved uranium is again coprecipitated using 10 mg ferric iron as the carrier and the precipitated ferric hydroxide (and uranium) is filtered. The precipitate is dissolved with HCl.

Procedure C⁵⁾: The solution is evaporated to dryness and is added 3 ml of conc HNO_3 . After evaporation to dryness is added 3 ml of nitric acid and again evaporated just to dryness. The residue is dissolved in 3 ml of 3N HNO_3 .

A cellulose column of 10 cm in length of the content is prepared and is washed thoroughly with 100 ml of ether-nitric acid solvent (ether containing 5% HNO_3). The solvent level in the column is adjusted until it coincides with the top of the cellulose. Sufficient cellulose pulp is added to the solution of the sample to absorb completely the aqueous nitric acid solution. The vial containing the absorbed sample is transferred to the top of the extraction column. The final traces of cellulose is transferred into the column with the solvent of ether-nitric acid not more than 10 ml. The pulp containing the sample is broken with a glass plunger and gently pressed down to form a continuation of the original column of cellulose.

Ether-nitric acid solvent is allowed to run out into a 200 ml beaker until the level of ether solution in the extraction tube reaches the top of the cellulose column. A further 10 ml of ether-nitric acid solvent is added to the top of the extraction tube and then the procedure with successive 10 ml portion of the solvent mixture is repeated until 100 ml of the eluate have been collected. A 20 ml of water is added to the eluate ether solution and the organic solvent is removed by evaporation. After the evaporation of the solution to dryness, 5 ml of conc. HCl is added and HNO_3 is removed. After evaporating to dryness, 5 ml of conc. HCl and 5 ml of water are added and traces of iron are removed by the ether extraction. A small amounts of conc HClO_4 and HNO_3 are added and is evaporated to dryness to decompose traces of organic substance present. The residue is dissolved by the addition of 0.2 ml of 1N HClO_4 and the concentration of uranium is determined polarographically by the MB2R method.

The procedures mentioned above were resulted after the fundamental preliminary examinations written below.

The separation of uranium from iron

Mixtures of 50 mg Fe and 10-50 μ g of UO_2^{2+} were treated following the Procedure C. Iron could be completely eliminated from uranium and the average of the recovery of uranium in this procedure was $97.1 \pm 1.5\%$.

The recovery of uranium in the coprecipitation process

The recovery of uranium in the coprecipitation with iron was investigated using artificial sea-water as medium. The results in the case in which Procedure B was omitted are shown in Table I. From Table I, the recovery in Procedures A and C can be regarded as almost 100%. It is also apparent that major elements in sea-water do not interfere the determination of uranium.

Table I. Recovery of Uranium from Artificial Sea-Water*.

U added (μ g)	U found (μ g)	Recovery (%)
20	19.8	99.0
30	31.1	103.5
30	30.3	101.0
50	49.6	99.3

* Corrected for the recovery in Procedure C.

The effect of minor elements

The effect of minor elements was tested. Table II shows the rough amounts of such ions in 20 liters of sea-water. Because the interference of such elements as As, P, Zr, Ce etc. can not be eliminated by the Procedure C alone, the Procedure B must be applied before the Procedure C. By the Procedures B and C, the interference of these elements could be eliminated. But the recovery through these processes was 72%.

Table II. Minor Elements Added*.

Metal	μ g added	Metal	μ g added
V(V)	1000	As(III)	200
Cu	4000	P	2000
Al	4000	Sn(IV)	120
Co(II)	200	Zr	10
Ni	100	Hg(II)	10
Mo(VI)	400	Ce(IV)	10
Cr(VI)	2	Th	10

* Equivalent to the content in 20 litres of sea-water.
Zr: imaginary value.

The determination of uranium in sea-water

The content of uranium in natural sea-water was determined by the procedures demonstrated in Fig. 2. The results obtained are shown in Table III. The content of uranium is 1.6 $\mu\text{g/L}$. This result seems to be reasonable comparing with that reported with the sample of the other parts of the ocean.

Table III. Uranium in Sea-Water.

Sample No.	U found ($\mu\text{g/L}$)
1	1.5
2	1.5
3	1.6
4	1.6
5	1.7
Average	1.58 ± 0.12

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The Relation between Chlorinity and Silicate Concentration of Waters observed in Some Estuaries V

Consideration upon the fitness of $[\text{Cl}]-[\text{SiO}_2]$ relation of the river water for the relation formula computed from the estuarine waters and upon the seasonal variation of the estimated constants

By

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Synopsis

This report is written for the purpose of clarifying whether we can represent $[\text{Cl}]-[\text{SiO}_2]$ relation of river water by expanding the relation formula estimated from the estuarine waters or not. And also some discussions on the variability of the estimated constants are included. Among 31 series of the observations in the estuarine region of the river Yorimo during the period from Feb. 19, 1955 to March 18, 1957, the relation, $A[\text{Cl}]+[\text{SiO}_2]=B$, can be admitted to each result of 27 series with considerable accuracy. So far as the present stage of this series of works is concerned, no essential difference of $[\text{Cl}]-[\text{SiO}_2]$ relation of the river water from that of the estuarine water can be recognized. The apparent seasonal variations of the constants are illustrated in Figs. 4-1 and 4-2. But because of the unstability of the estimated constants, plentiful consideration should be given to the variability and the magnitude of fluctuation of the estimated constants, before any discussion on the variation of the constants due to any factor is given.

The senior author and his coworkers (MAÉDA 1952, 1953, MAKIMOTO *et al.* 1955 and MAÉDA & TSUKAMOTO 1959) had studied on the relation between chlorinity (dominant in sea water) and silicate concentration (dominant in river water) of the waters observed in the estuaries and found that they fit for the formula, $A[\text{Cl}]+[\text{SiO}_2]=B\cdots(1)$, with considerable accuracy and also **B** of the above-mentioned formula estimated from many estuaries in Japan in various seasons bears the following relation to **A** with considerable accuracy:

$B=0.755+17.07 A \cdots (2)$. Moreover they reported that these constants seemed to take the values somewhat specific to the waters pouring into them, although some seasonal and occasional variations of the estimated constants are suspected from the results.

The chlorinity of the river water is able to be regarded nearly as equal as zero. Therefore, substitute zero for the term $[Cl]$ of formula (1), then the relation, $[SiO_2]=B$, is inducible, which means that one of the constants, **B**, indicates the silicate concentration of the river water. Thus the silicate concentrations of the estuarine waters at respective chlorinities are thought to be depended on the silicate concentration of the river water flowing into it. Here it is easily supposable that no phenomenon other than simple dilution or simple mixing with the sea water occurs to the silicate contained in the river water after it is flowed into sea, furthermore the silicate concentration of the river water is regulated by the histories of the water in land—for examples, nature of soil through which the water passed, manner of flowing, current velocity, etc. But the above-mentioned relations are estimated from the waters, the chlorinity of which is from 5‰ to 19‰. Therefore, a question, whether or not the relation between chlorinity and silicate concentration observed in the river water actually fit for the formula estimated from the estuarine waters, should be examined, for the purpose of finding out the lowest limit of chlorinity applicable to the formula—in other words, whether **B** really represents the silicate concentration of the river water or not. The seasonal variation of the estimated constants should also be examined, before the factors controlling the constants are studied, as a preliminary step of the study. Some results of the examinations confirming the above-mentioned assumptions on the data obtained from the estuarine and river water of the river Yorimo during the period from Feb. 19, 1955 to March 18, 1957 will be illustrated in this report.

Before entering into the subject, we must express here our sincere thanks to Prof. Dr. D. MIYADI of the Zoological Institute of Kyoto University and Dr. T. TOKIOKA of the Seto Marine Biological Laboratory of the same university and Prof. Dr. I. MATSUI of the Shimonoseki College of Fisheries, for their valuable advices and criticisms given on the present series of works. And also we must record here our hearty thanks to Messrs. M. KITANI, K. OGATA and any other staffs of the Takada Branch of the Oita Prefectural Fisheries Station, who helped us much during this study.

Methods

Waters analyzed were sampled by KITAHARA'S bottle from the layer a meter above the bottom at each station represented by filled circle and white one in Fig. 1 at flood-tide in every other spring tide during the period from



Fig. 1. Sketch chart of the area surveyed. Note: For the stations represented by filled circle and white one, see "methods"; for St. I and St. II, the first several lines in "consideration".

Feb. 19, 1955 to March 18, 1957. But during the latter half of the period, from April 12, '56 to March 18, '57, the stations represented by while circle were thinned out, against the fact that the samples collected at ebb-tide were added; here when it is ebb-tide and the station can not attain to a meter deep, the surface water is sampled in place of a meter one.

And the chlorinities and silicate concentrations of the water samples thus collected were determined respectively by MORH's silver titration method and DIÉNERT-WANDENBULCKE's colorimetric method just after they were collected. After the same preliminary consideration to that mentioned in the fourth report of this series (MAÉDA and TSUKAMOTO 1959) were given, the data were treated statistically by the same method as described in the second report (MAÉDA 1953).

Results

The $[Cl]-[SiO_2]$ relation formulae estimated from the results of respective observations and their significance levels were shown in Table 1. The chlorinity and silicate concentration of the water samples collected at every observation at flood-tide during the period from Feb. 19, '55 to March 18, '56 well fit for the formula, $A[Cl]+[SiO_2]=B$. But among the examples obtained at respective observations at flood-tide during the period from April 12, '56 to March 18, '57, there are some examples the $[Cl]-[SiO_2]$ relation of which can not be regarded fitting for this formula, because of negligibly small deviations of chlorinity and of silicate concentration caused by thinning out the stations. Contrary to the results of the observations pursued at flood-tide, the same results as the observations done at flood-tide of the preceding years can be admitted to most of the observations done at ebb-tide of this period, because of severe deviations of chlorinity and of silicate concentration due to diversity of the influence of the river water on respective stations.

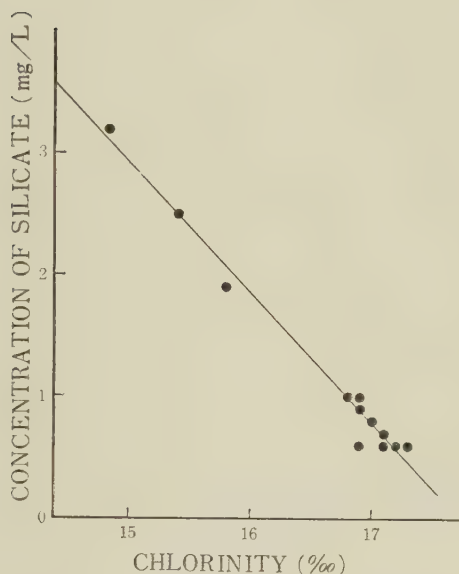


Fig. 2-1. An example of silicate-chlorinity relation graph (Flood-tide, April 26, '55).

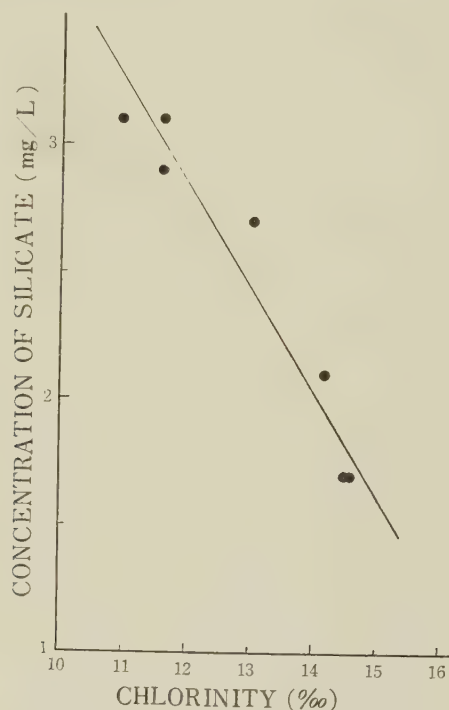


Fig. 2-2. An example of silicate-chlorinity relation graph (Ebb-tide, Jan. 17, '57).

Consideration

1. Examination into fitness of the $[\text{Cl}]-[\text{SiO}_2]$ relation observed in the river water for the relation formula

The river waters sampled from St. I (from August 1955 to March 1957) and St. II (from August '55 to April '56) during the same observation when samples of the estuarine waters at flood-tide were collected, for the purpose of examining the fitness of the $[\text{Cl}]-[\text{SiO}_2]$ relation observed in the river water for the relation formula. But unfortunately, the results of the analyses tell us that they are not the purely river waters but are more or less affected by the sea water. Consequently, even these samples can not be the samples strictly suitable for this purpose, while it is thought to have a little value worthy to be examined on the question, whether these samples are able to be regarded fitting for the corresponding formulae or not, because their chlorinities were far lower and their silicate concentrations were far higher than those of any estuarine samples collected during the same observations. Accordingly, whether or not the $[\text{Cl}]-[\text{SiO}_2]$ relation of these samples can be regarded to follow the corresponding formulae are examined through the fol-

lowing formula, in which it is set as null hypothesis that the relation well fits for the relation formula :

$$F_0 = \frac{\{y_0 - \bar{y} - a \frac{x_0 - \bar{x}}{S_x}\}^2}{N-2 \left\{ 1 + \frac{1}{N} + \frac{(x_0 - \bar{x})^2}{S_x} \right\}}$$

Here, degrees of freedom of F , n_1 and n_2 , are respectively one and $(N-2)$; $\bar{x}$ and $\bar{y}$ are respectively mean of chlorinity and that of silicate concentration; a is the regression coefficient of the relation formula, $S_x = \sum_{i=1}^N (x_i - \bar{x})^2$; $\Delta = S_y - \frac{C^2}{S_x}$, in which $S_y = \sum_{i=1}^N (y_i - \bar{y})^2$; while $C = \sum_{i=1}^N (x_i - \bar{x})(y_i - \bar{y})$; N is the size of sample, all of them are estimated from the group of samples from which the relation formula is estimated. And x_0 and y_0 are the chlorinity and silicate concentration of the sample the fitness of which is going to be examined.

The results of this series of examinations were also illustrated in Table 1, in which the computed F from the data obtained at St. I and St. II were respectively represented by F_I and F_{II} . These results reveal that the most of $[\text{Cl}]-[\text{SiO}_2]$ relations of the river waters are unable to regard as out of the corresponding formula. Moreover, examining the samples of the river water the $[\text{Cl}]-[\text{SiO}_2]$ relation of which can be regarded as out of the relation formulae—although such examples are not so frequently observable—, we are aware that even most of the $[\text{Cl}]-[\text{SiO}_2]$ relations of them are not so far from those represented by the corresponding formulae, but they can not be regarded as fitting for the relation formulae because of such reason as Δ is extremely small comparing with S_y , which means that the silicate concentrations of the estuarine waters change strictly depending on their chlorinities consequently confidence limits of silicate concentration at respective chlorinities are very narrow. Therefore, generally speaking, so far as the present stage of this study is concerned, it may be said safely that any essential difference of $[\text{Cl}]-[\text{SiO}_2]$ relation of river water from that of estuarine waters can scarcely be recognized.

2. Seasonal variation of the estimated constants of relation formulae

The results in the preceding reports of this series of works reveal that **B** of relation formula changes depending on **A** keeping the following relation: $B = 0.755 + 17.07 A$. So, for the purpose of proceeding the further consideration easily, whether or not the similar relation can also be admitted between the series of constants of the relation formulae newly estimated is examined. Fig. 3 tells us, without requiring any statistical examination, that it is sufficient to discuss upon a single series of constants. Accordingly, series of **A** is taken up hereafter, and their seasonal variations are shown in Figs. 4-1 and 4-2. We are aware from these figures that considerably severe seasonal

Table 1. Constants of the $[Cl]-[SiO_2]$ relation formulae, their significance levels and the results of the examination on fitness of the relation of the river water to the relation formulae.

Date	Tide level	N	Retation formula			Fitness of river water		
			A	B	F ₀	Level of significance	F _I	Level of significance
Feb. 19, '55	Flood	12	0.743	14.805	39.19	<0.001		
March 27	"	Omitted, because no silicate concentration except that at St. 7 was measured.						
Apr. 26	"	12	1.069	18.945	505.72	<0.001		
May 23	"	12	2.983	50.785	37.00	<0.001		
June 23	"	12	0.886	15.556	68.61	<0.001		
July 21	"	12	0.979	15.724	135.37	<0.001		
Aug. 22	"	12	1.416	19.411	30.42	<0.001	0.239	0.05<
Sept. 20	"	12	0.401	7.318	12.61	0.001~0.01		
Oct. 22	"	22	1.828	32.193	72.27	<0.001	11.698	0.001~0.01
(Except St. 1)								
Nov. 18	"	11	0.859	15.426	9.78	0.01~0.05	0.746	0.05<
"	"	12	1.718	29.948	26.09	<0.001	2.544	0.05<
Dec. 15	"	Omitted, because no chlorinity except that at Sts. 1, 4 and 10 was measured.						
Jan. 16, '56	"	12	1.127	20.663	24.13	<0.001	0.851	0.05<
Feb. 17	"	9	1.616	29.729	290.73	<0.001	4.201	0.05<
(Except Sts. 1&2)								
Mar. 18	"	7	1.191	22.266	11.81	0.01~0.05	0.364	0.05<
"	"	10	0.881	15.737	612.408	<0.001		
Apr. 12	"	7	0.795	14.747	8.33	0.01~0.05	0.137	0.05<
"	Ebb	7	0.755	15.766	109.41	<0.001	0.266	0.05<
May 11	Flood	7	0.731	14.005	22.91	0.001~0.01	9.59	0.01~0.05
"	Ebb	Omitted, because it was rain.						
June 11	Flood	7	0.850	15.090	944.59	<0.001	119.848	<0.001
"	Ebb	6	0.572	14.735	51.83	0.001~0.01	9.059	0.01~0.05
July 11	Flood	7	—	—	1.29	0.05<		
"	Ebb	7	0.785	14.016	24.97	0.001~0.01	0.005	0.05<
Aug. 13	Flood	6	—	—	1.91	0.05<		
(Except St. 5)								
Sept. 25	Ebb	5	1.316	21.523	121.39	0.001~0.01	20.026	0.01~0.05
"	Flood	6	0.604	10.641	42.15	0.001~0.01	0.520	0.05<
"	Ebb	Too low chlorinity and omitted.						
Oct. 22	Flood	7	—	—	0.06	0.05<		
"	Ebb	7	0.857	17.784	23.85	0.001~0.01	6.298	0.05<
Nov. 5	Flood	Omitted, because no chlorinity was measured.						
"	Ebb	7	0.747	16.223	36.86	0.001~0.01		
Dec. 10	Flood	7	—	—	0.14	0.05<		
"	Ebb	7	0.744	12.238	67.67	<0.001	1.807	0.05<
Jan. 17, '57	Flood	Omitted, because silicate concentration was too low to measure.						
"	Ebb	7	0.394	7.546	62.93	<0.001	0.320	0.05<
Feb. 13	Flood	Omitted, because silicate concentration was too low to measure.						
"	Ebb	7	0.795	13.739	32.92	0.001~0.01	0.05	0.05<
Mar. 18	Flood	7	0.384	7.165	9.19	0.01~0.05	20.182	0.001~0.01
"	Ebb	7	0.846	14.922	244.29	<0.001	3.114	0.05<

Notes: Relation formula: $A[Cl]+[SiO_2]=B$, $[Cl]: \%$, $[SiO_2]: mg/L$

N	N-2	$F^{1N-2}(\alpha)$	
		0.001	0.01
5	3	167.5	34.12
6	4	74.14	21.20
7	5	47.04	16.26
9	7	29.22	12.25
10	8	25.42	11.26
11	9	22.86	10.56
12	10	21.04	10.04
			0.05
			10.13
			7.71
			6.61
			5.59
			5.32
			5.12
			4.96

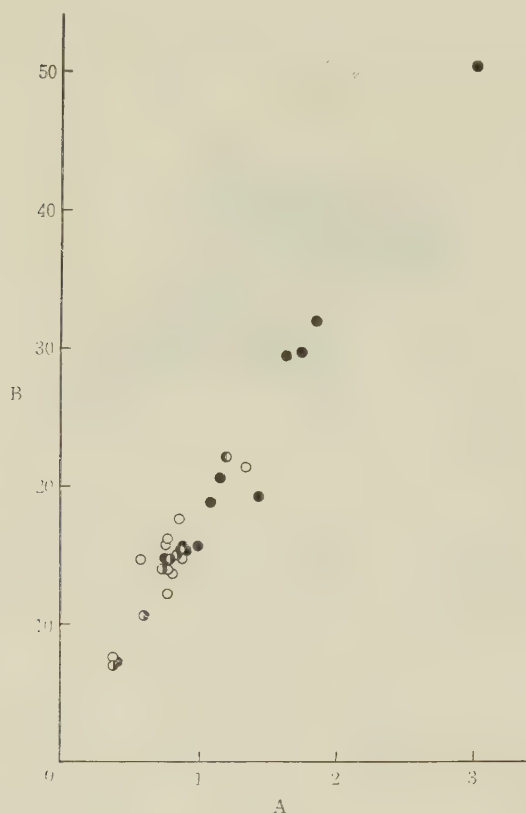


Fig. 3. Relation graph of the constants, A and B, of the
silicate-chlorinity relation formulae.

Note: ●: Flood-tide, Feb. '55—March '56 (all station)
◐: Flood-tide, Feb. '55—March '56 (some are rejected)
◑: Flood-tide, April '56—March '57
○: Ebb-tide, April '56—March '57

variations of constants are observable moreover there are considerably conspicuous differences of the features of seasonal variations of the relation formulae between those obtained during 1955—'56 and those during 1956—'57. But, here, we must call attention to the fact that Fig. 4-2 is drawn on the relation formulae at ebb-tide against the fact that Fig. 4-1 is constituted from the relation formulae at ebb-tide. This fact may more or less contribute to causing the above-mentioned differences; while, here fortunately we have some pairs of relation formulae each one of which is obtained at flood-tide while the other is at ebb-tide occurred some six hours later than the former. And they are respectively 0.795 (flood) and 0.755 (ebb) on April 16, 1956, 0.850 and 0.572 on June 11, and 0.384 and 0.846 on March 18, '57. Thus, the constants are very unstable and we can hardly find out any definite tendency of

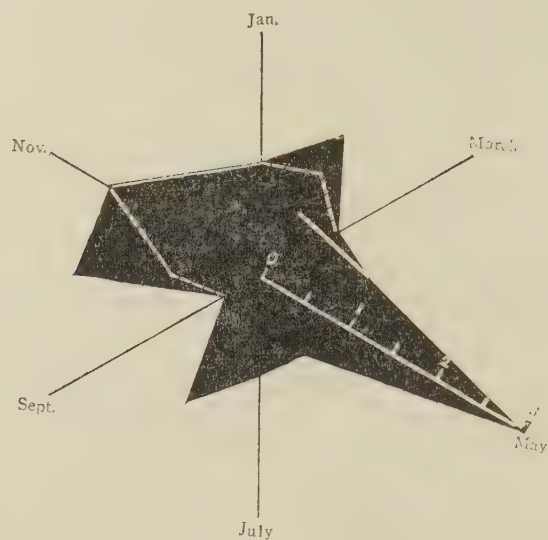


Fig. 4-1. Seasonal variation of a series of constant, A , of relation formulae (Flood-tide, Feb. '55—March '56).

Note: $\square$: Lacking the observation.
 $\square\square$: Some stations were rejected.

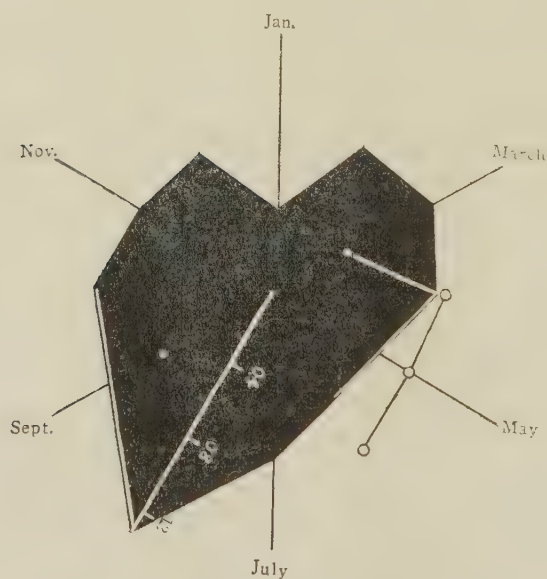


Fig. 4-2. Seasonal variation of a series of constant, A , of relation formulae (Ebb-tide, April '56—March '57).

Note: See the note under Fig. 4-1.
 Circle indicates the value at flood-tide.

change—whether keeping invariable or not moreover whether increased or decreased. Accordingly, no definite discussion can be given on the seasonal variation of the relation formulae.

Conclusion

The facts clarified in this report summarized, the below-mentioned facts are safely said as the conclusion of this report: 1) There may be no need to pay any attention, so far as the chlorinity and silicate concentration are concerned, to the possibility of occurrence to any other phenomenon than simple mixing with water or phenomenon the effect of which is proportional to it. 2) Because of the unstability of the estimated constants, no discussion can be given on the variation of the constants due to any factors, including regional specificity and seasonal variation, without any detailed consideration upon the variability and magnitude of the fluctuation of the constants, although significance levels of relation formulae and of the **A—B** relation are so high enough as to admit of no doubt about their existence.

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Uses of Radio-isotopes for Tracing Predations of Trout-Fry

By

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Introduction

During recent several years, the use of the radio-isotopes has been remarkably developed in many aspects of scientific researches, and the possibility of the application of this to the field of fisheries biology or of marine biology has been thought by some scientists, including the author himself.

As the radio-isotopes conveniently label organic and inorganic materials for tracing, these have been used for the research of the metabolisms of the aquatic organisms either in case of direct material up-take from the environmental water or in case of food in-take.

Beside this type of use of radio-isotopes in marine biology, the use of this for the marking experiment was planned and tried in the laboratory tank by the author and ICHIKAWA¹⁾ in 1955. Co⁶⁰ pin of 100 μ c was enveloped into a glass tube of 5 mm long and of 1 mm diameter and used as γ ray radiation source. The three locations in gold fish body (about 10 cm long) were selected to insert this source for comparison, i.e. (1) inside of abdominal cavity under pectoral fin, (2) inserted through anus to abdominal cavity, (3) inserted into dorsal muscle. By rearing these treated fish about one year long, this type of marking is found to be practicable except in case of dorsal muscle insertion as the source had gone off from fish during the experiment. The existence of this γ ray source in living fish can be ascertained by the usual Geiger-Muller Survey Meter easily from the distance of about 1 meter, and by careful measure from 2 meter distance in air and from 39 cm deep in water. This laboratory experiment suggests the possibility to apply this method to some field experiment if the recovery is expected to be done completely such as in case to liberate the marked fish into some small pond or to a stream.

On the other hand, the results of the radiological surveys on the Pacific tunas contaminated by radio-active materials born by nuclear bomb test explosion since 1954 suggest the possibility of use of radio-isotope as the experimental tracer, if the amount is small enough and the distribution of these is only in a restricted area, to prevent the possible public hazard to know both on the migration of fish from the point of contamination²⁾ and on the

metabolic channels of these materials from sea-water to fish through food organisms.³⁾

As these previous experiments and surveys suggest, the authors, by the cooperation of Mr. ONODERA and Mr. KODAMA of Fresh-water Fisheries Research Laboratory, and of Mr. SUZUKI of Kanagawa Prefectural Fisheries Laboratory, besides of Mr. NOSE, Mr. SHIMIZU and Miss MATSUBARA of our Institute, tried to use this in a restricted field as mentioned below to know about the predation of rainbow trout fry liberated to that area based upon the fact that radio-isotope in rainbow trout fry shall go to the predator after the prey had taken by these.

Field of Experiment

A mountain stream, called Shiokawa, a branch of the river Nakatsugawa, in Kanagawa Prefecture, had been used by Fresh-water Fisheries Research Laboratory and Kanagawa Prefectural Fisheries Laboratory since several years ago for the experimental liberation of trout fry and ecological research on mortality rate.⁴⁾ The stream runs inside of steepy V-shaped valley, and thought to be easy to prevent fishing by intruder. However, in several spots notice plates to prohibit the fishing and to announce that radio-isotope is used were settled. Moreover, by the understanding and cooperation of the neighbouring village people, the area was kept free from intrusion of fisherman.

Having a high water-fall at upper stream end, and a fence and a trap at lower stream end, the liberated fishes had been kept confined within the experiment area, 330 m long and 3 to 5 m wide.

For the first year, 1959, entire area was used, liberating fry at the upper stream end, the foot of the fall. But in 1960, the liberation was done at middle point, making a small dam above it, and only the lower half of the area was used.

Zn⁶⁵ injection method

For the first year's experiment, 1959, Zn⁶⁵, having physical half life as 250 days, was selected for the radio-isotope to be injected to rainbow trout fry as the tracer by the following reasons. (1) the biological half life is adequate for the length of the period of research, (2) the maximum permissible amount for human food intake is relatively large, (3) γ emitter and easy to detect from long distance, and (4) the distribution in fish body is rather homogeneous.

The amount of Zn⁶⁵ to be injected to fish fry was decided as 0.03 cc of 30 μ c/cc solution of Zn⁶⁵Cl, i.e. 1 μ c, per individual of trout fry, having average body weight 1.2 g in the small size group and 2.7 in the large size group. Though the more the amount of the radio-isotope is injected, the more easy to trace and to detected, this amount is a limit to keep human intake as 1/100

of the maximum permissible level in case one fish in a day was taken by a man continuously. The concentration of river water, if all of these go out to water, is calculated as 10^{-4} of the maximum permissible concentration of drinking water, in the worst condition.

However, the radio-activity from this amount in fish body had been well recognized from outside of the body in closer examination by usual Geiger Muller survey Meter, during two or three months after the injection and liberation. Fig. 1 shows the biological decay of Zn^{65} in trout fry, obtained by the radio-assay of ash of fish, sampled from the stream after the injection. This shows, even in days 5 months after the liberation, if the fish can be ashed and assayed, it is easy to detect the radio-activity.

The effects of the radiation from the internally kept Zn^{65} upon the survival and growth of the trout fry was checked by frequent sampling from the stream. Fig. 2 shows the difference of survival rate between the treated

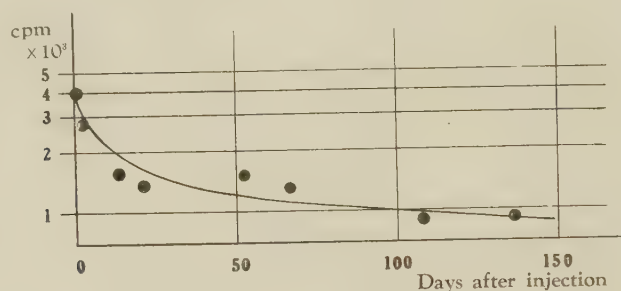


Fig. 1. Biological Decay of Zn^{65} in a Rain-bow Trout Fry.

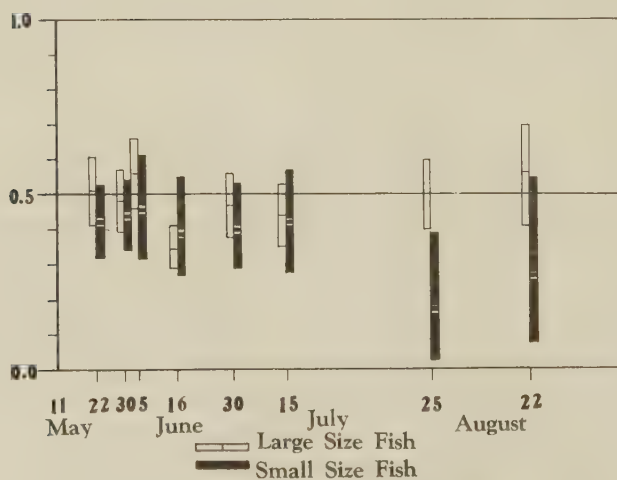


Fig. 2. Ratio of Radio-active Fish to Total Number of Fish in Test Fishing.

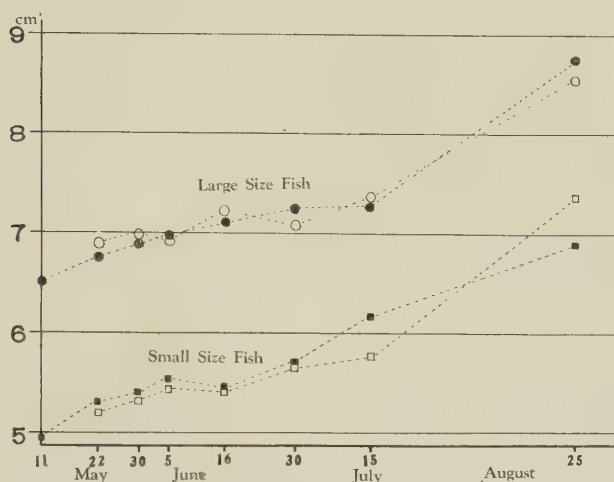


Fig 3. Comparison of Growth Between injected (black) and non-injected (white) Rain-low Trout Fry.

and non-treated fish is negligible except in case of the small size group. Fig. 3 shows the difference of the growth between the two groups, treated and non-treated, but was not significant.

Result of Survey

Since after the liberation, 11th of May, 1959, various sorts of materials in the vicinity were monitored around closely by G-M Survey Meter, and many plants and animals, as well as soil, water and others, in and around the stream, were collected for closer examination by G-M counter. The results of the surveys done frequently during the research period enlightened the presence of radioactivity in only three kinds of materials among so many examined.

The first was the excrete of a bird, which later found to be of *Cinclus pallasii*, being a small white patch spread on the surface of rocks around the stream. A couple of this bird nested on mid-cliff of water fall at the point of liberation, and breeding at the time of liberation. This excrete was found to be radio-active always and collected for exact radio-assay.

The second was the adult trout which had liberated last year as fry and grown up to the adult of larger than 15cm long. 11 individuals of these were caught occasionally by test fishing up to 7th of July, and all of them were found to be radio-active. These fishes were asked for the exact radio-assay.

The third was a firesh-water crab, *Eliocheir japonicus* caught alive in the in the stream, and found radio-active by radio-assay. However, about the possibility that the crab can take the living trout fry, Mr. ONODERA thought he

does not take the living one after a rearing experiment in the tank with living trout fry in it, but may take after he died as usually the scavenger does.

By this radiological survey, the kind of the predator of the trout fry liberated in this stream was identified as a bird and adult trout, however, the author hoped to know further about the amount of trout fry taken by these predators during certain period. For this purpose, the following calculation methods were established by using the data obtained by some additional laboratory experiments to interpret the data obtained from the field survey.

Methods to know the Quantity of Predation

(1) **Predation by adult trout** By knowing the radio-activity of Zn^{65} in the adult trout, to obtain the number of the radioactive trout fry taken by him, it is necessary to get a proper ratio of intake (f_1) about Zn^{65} through food (trout fry) to adult fish, by some laboratory experiment.

Thus, several trouts of the same size as found in the stream were reared in the tank, and fed by the fish meal jelly made of the Zn^{65} injected trout fry by known amount of radioactivity. These trouts were sacrificed after 20 hours (time decided enough to complete digestion, absorption and excretion by experiments), and radio-assayed. Then, as the ratio of radioactivity in the fish to that in the food taken, $f_1=0.24$ was got as the average of experimented individuals. Though the radioactivity of Zn^{65} in the adult trout is reduced as time passed since the intake by physical decay and by excretion, however this trend of the biological decay of Zn^{65} in the adult can not be different from that in the fry. Therefore, when the predation occurred is not the matter to

Table 1. Predation of Rain-bow Trout Fry by Adult Trouts.

Large Trouts	Date caught	Locality caught	x_i cpm (Zn^{65} activity in Adult Trout)	y individual (Number of Fry Taken by Adult Trout)
Rain-bow Trout	May 22	B-10	570	2.4
Mountain Trout	May 30	A-10	80	0.4
Mountain Trout	May 30	A-10	273	1.4
Rain-bow Trout	May 30	A- 7	683	3.5
Rain-bow Trout	May 30	B- 4	238	1.2
Rain-bow Trout	June 5	A- 6	88	0.5
Mountain Trout	June 16	A- 8	1,176	7.3
Mountain Trout	June 16	A-10	196	1.2
Rain-bow Trout	July 7	C-13	53	0.4
Rain-bow Trout	July 7	C- 8	70	0.5
Rain-bow Trout	July 7	A- 6	2,120	14.7
11 Individuals				33.5

consider to calculate the number of trout fry taken by that fish from the radioactivity remained in it.

By this reason, a simple formula as follows can be established to calculate the number of trout fry taken by an adult trout up to the time when sacrificed (y).

$$y = \frac{x_i}{f_1} \times \frac{2}{a_i}$$

Where x_i is the radioactivity in the adult trout body when sacrificed and a_i is the radioactivity in a single trout fry at that time. The reason why 2 is here is that the number of the Zn^{65} injected fry was just a half of the trout fry liberated to the stream.

y calculated and x_i used for it about each individual of the predator were listed in Table 1 here.

(2) **Predation by bird** Being a wild bird, as *Cinclus pallasii* is hard to keep alive in captivity for the laboratory experiment on metabolism, a domestic pigeon was used as the substitute in the experiment to obtain the ratio of the excrete radioactivity to that had taken as food, trout fry, about Zn^{65} metabolism (f_2). During the experiment the known amount of the Zn^{65} injected trout fry meal was given daily to pigeons, and the radioactivity in the daily excrete was radio-assayed. Then dividing the latter by the former, $f_2=0.84$ was obtained.

In this case, beside figure of f_2 , it is necessary for the calculation of the number of trout fry taken by a bird in day to get an adequate figure for the frequency of excrete in a day. The figure for it got by the pigeon rearing experiment would not be proper, as it may differ from that of wild bird. And, for us it was difficult to get it by field observation and survey. Very fortunate for us that Mr. Tadakazu ISHIHARA of Kyoto University has been studied the ecology of this kind of bird, and gave us his information based upon the observation that a couple of this bird excrete 126 times a day (N).

As for the average radioactivity of a single excrete collected from the surface of rocks around the stream (x), 100 cpm was obtained from 5 samples collected 5th of June (5 to 240 cpm).

By these data, the number of trout fry taken by a couple of the bird nested there in a day (y) can be obtained by the following formula.

$$y = \frac{Nx}{f_2} \times \frac{2}{a_i}$$

Taking a_i as the radioactivity of trout fry at the same day when the excrete was collected (1,500 cpm at June 5th), y can be as follows.

$$y = \frac{126 \times 100}{0.84} \times \frac{2}{1,500} = 20$$

This means that a couple of this bird was taking 20 trout fry every day

at that time.

Interpretation of the Results of Calculation

On the other hand, Mr. ONODERA and others were working on the survival rate of the liberate trout fry by frequent test fishing in that stream simultaneously. Their results of survey are not published yet, however, by their courtesy, getting informations on their findings, the author can interpret our results by using these.

On May 11th, 1959, 2034 individuals of rainbow trout fry was liberated at the foot of water-fall in upper end of the stream, and among them just a half was injected by Zn^{65} . At July 7th, 57 days later than the liberation, Mr. ONODERA and others calculated the number of the trout fry survived in the stream as 782 individuals. In other words, $2034 - 782 = 1252$ individuals were lost during this period.

Then the number of trout fry taken by 11 adult trouts during the same period was calculated as 33.5 individuals as Table 1 shows.

Beside this, if the size of daily predation by a couple of this bird nested near the liberation point could be the same all through the period (57 days) as that of June 5th, the total number of trout fry lost by this type of predation is $20 \times 57 = 1140$ individuals. Then, putting the both altogether, the total number of the victims is $33.5 + 1140 = 1173$. As the number of individuals lost during the period was calculated as 1252, then the difference is $1252 - 1173 = 79$. The difference may be logically of natural mortality, but the range of the accuracy of the parameters used this calculation is not enough to conclude this precisely, especially of x .

This calculation says that during almost two months the trout fry liberated in this stream lost 61.5% of total number, and only 38.5% of it survived. Among this loss, 94% of it was caused by the predation of animals, and 97% of the predation was by a couple of this bird and 3% of it by the adult trouts survived in the stream.

Co^{60} Pin Method

Zn^{65} injection method applied to the trout fry liberation experiment of 1959 brought the knowledge on the size of the predation more than expected at the beginning. However, this method requires the close examination of radioactivity about the samples thought to be suspicious and the samples able to be collected for the examination. Therefore, there are some possibilities of existence of some predators other than the two already known in the 1959 experiment, being unsuspected or uncollected.

Thus from May 1st 1960 on, another type of radio-isotope marking to trace the predator of trout fry was planned in the same stream as was used

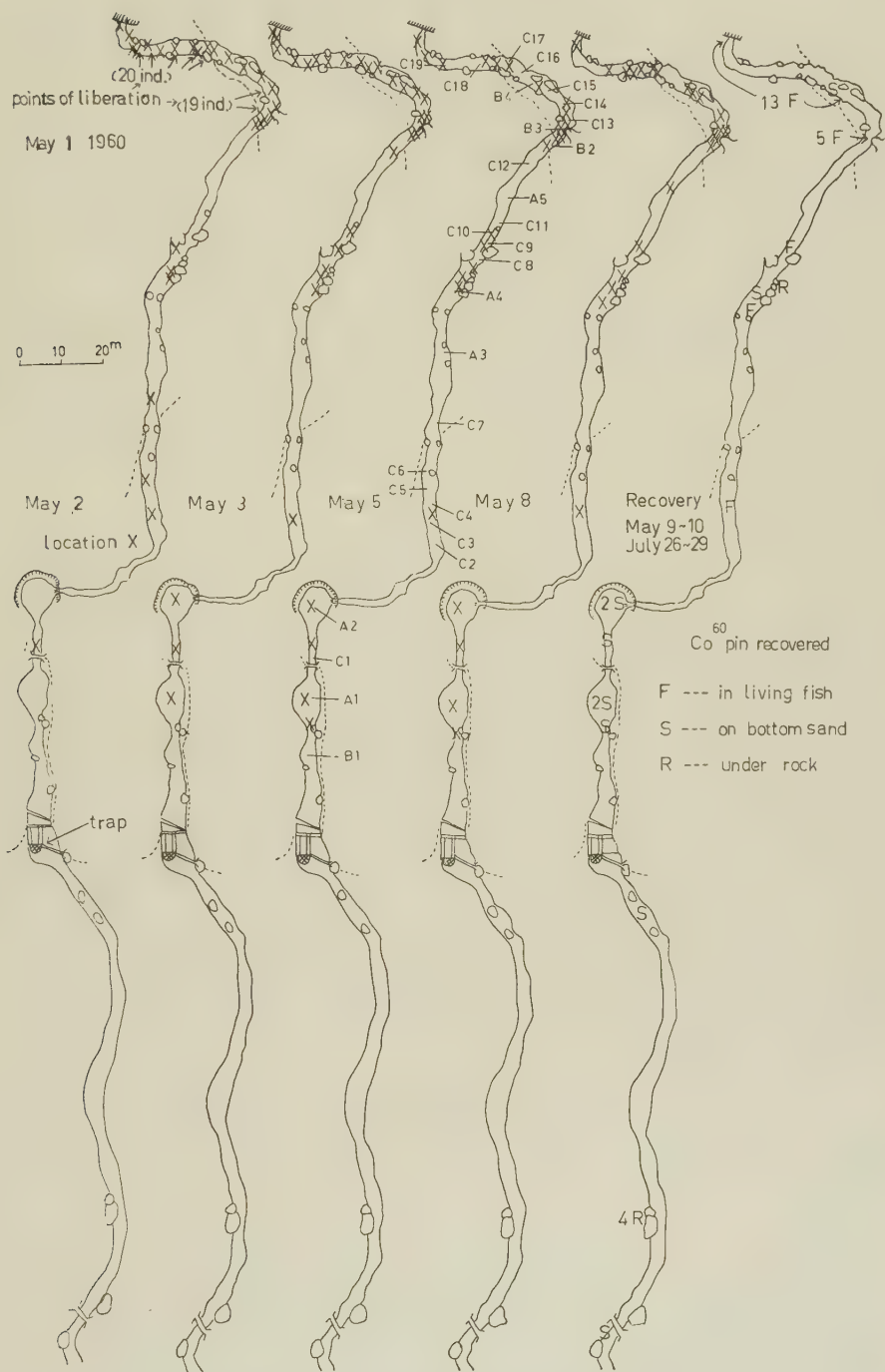


Fig. 4. The Points where the Rainbow Trout Fry liberated (→), the Locations where the Radiation Sources were detected (X) and the Locations where the Co⁶⁰ Pins were recovered (F, S, R).

in the last year, but only the lower half of the stream was used making a dam near the middle point.

Instead of injecting the radio-isotope to fish body, a Co^{60} pin for medical treatment was cutted into a small piece, about 1 mm long and 0.8 mm diameter, and either coated by paint or enveloped into plastic capsule. This was inserted into abdominal cavity of trout fry through 1 mm wide cleft made by a tip of sharp knife. One piece of radiation source, about $100\mu\text{c}$ for each, was in one individual of fish, which has average length as about 9 cm and average body weight as about 8 g (3.5 to 9 g).

At noon of May 1st, 1950, 191 unmarked and 39 Co^{60} marked rainbow trout fry were mixed and liberated to the points of two parts in the stream as shown in map in Fig. 4. Since that afternoon, the vicinity around the stream was monitor at least once in a day, and some of the results of monitoring were shown in map in Fig. 4.

Tracing the Distribution

The existence of Co^{60} pin either inside or outside of fish body was easily detected from the distance by Crystal Scintillation Survey Meter ($1\frac{1}{2}$ " NaI crystal head) and by G-M Survey meter. Especially the former is so effective that the source could be recognized even from 30 meter distance in air and 1 meter deep in water. However, as this instrument is so sensitive that it was difficult to count the exact number of radiation sources when they located close together. In case of under-water survey, to recognize the two sources of radiation as two, it was necessary to keep them apart more than one meter from each other. So often, in the experiment, the radiation sources were found beneath huge sized rocks, as these were excreted there by the predator. In this case, as the shielding effect of rock against radiation is so prominent, it was necessary to monitor from various direction around the rock to detect the radiation source, and was difficult to count the exact number of sources under the rock.

However, as the instrument was more sensitive and the radiation dose was more than the previous experiment, almost of all the existense of radiation sources was detected by the survey of every day as the locations of finding were marked in map in Fig. 4.

In this experiment, as no test fishing was done until 9th of May when we started to recover the radiation source, so it is difficult to know either this radiation source was inside of fish or outside being so by predation and excretion, especially about the findings in later days. However, supposing no predation had happend yet in the time shortly after the liberation, the locations of findings at 3 hours and one day after the liberation show fairly fine pattern of redistribution happened right after the liberation. It was reco-

gnized that a few fish went upstreamward about 10 meter distant, and a few went downward about 150 meter distant from the point of liberation within one day.

As the day goes on, the wider the distribution was found. But, some of the spots in extremity were found to be already excreted by the predator when recovered.

The locations, where the radiation source was detected, in maps in Fig. 4 were rather fixed in every survey during several days. Probably this was caused partly by the character of fish to prefer to be in some peculiar point, and partly by the fact that the radiation source was already on the bottom sand as it was found to be so by the recovery.

Recovery of Co^{60} Pins

As we hoped to know about the state of the radiation sources by closer examination, and also for the safety, we tried to collect all the radiation sources we could detect since 9th of May.

The method to collect the radiation sources from the area was as follows.

Firstly we monitored the area thoughly, and recorded the points where the radiation was detected. When the source was found under water, with a suspect that might be a living fish, an electric shocker (100 V D.C.) was applied to that spot to drive the fish out and catch it by a dip net. This method was proved to be effective in many cases, and the paralised or semiparalised fish was easily caught. However, in some cases, inspite of frequent trial of the electric shocking, no fish comes out and the radiation source remained still without showing any indication of moving. In this case, the bottom sand or any other suspicious materials were collected for close examination, until the radiation source, Co^{60} pin, comes out and no radiation was remaining there.

This kind of effort of collection in 9th and 11th of may brought the following recovery. (1) 21 individuals of living fish keeping Co^{60} pin inside of body. (39 individuals were liberated at May 1st.) (2) 4 pieces of Co^{60} pin were found from the sand in the bottom of the stream.

However, there was a depth, named A-1, and though we recognized the presence of the radiation source or sources at the bottom, but we could not get these. Beside this, in some spots on beach or on land we could detect the radiation, but could not collect these as these were under huge sized rocks. But, all the living fishes were thought to be recovered at this time.

To recover the remaining sources, again on 26th to 29th of July, equipped with more powerful instruments to dig these out (boat, ropes, scoop, magnet and others), the effort was applied to the spots already recorded, and taken out all the detected radiation sources from the vicinity.

By this effort, the following recovery was added to the previous recovery.

(1) 6 Co^{60} pins found on the bottom sand. (2) 5 Co^{60} pins found from two spots in a hole under huge rocks.

The total recovery from 39 individuals of fish bearing Co^{60} inside and liberated to the stream at May 1st is as follows.

- (1) 21 individuals of living trout fry.
- (2) 10 pieces of pins found on the bottom sand.
- (3) 5 pieces of pins found in a hole under rocks.

These altogether make 36, and 3 pieces were missed in spite of the tremendous effort of monitoring and collection.

Predation Supposed from the Recovery

About the cause of the existence of Co^{60} pins on the bottom sand, as the pins inserted into the abdominal cavity of trout fry hardly goes out unless the fish die and decay—known by the experiment¹⁾—, it is quite logical to interpret this as the result of the predation of some animal and pins were excreted as a whole to the stream.

About the kind of animal of the predator, as pins were excreted in the water, and also by the experience of the last year, the adult trouts, which were also found plenty in this stream in this year, would be reasonably suspected with strong possibility. If it really so, the adult trouts had taken about 1/4 of the liberated trout fry within 8 days, and is the most effective predator in this year.

About the cause of the existence of Co^{60} pins in a hole under huge rock, we can not think any cause other than to be the result of predation and excretion by *Chimarrogale platycephala*, a Japanese Water Shrew, a rat or mole shaped mammal, which was found abundant around there and one of them was caught by electric shocker and net and identified to be this. The hole where Co^{60} pins were found looks like an elaborated nest, as shown in Fig. 5, made of tunnels about 3 cm diameter, and having two entrances facing to the stream which located about a half meter above the usual water surface. These tunnels were found either between or under huge rocks which was hard to

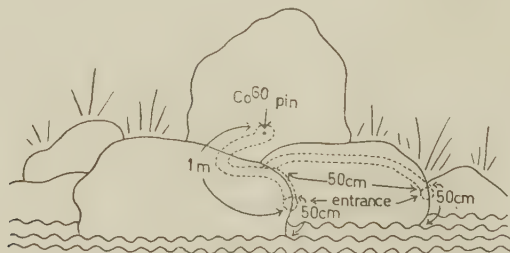


Fig. 5. A Nest supposed to be of Water shrew where Co^{60} pins were found.

to move by man power. The point where Co^{60} was found and supposed to be the spot to excrete was under a biggest rock. Fig. 5 was drawn about the one found at C-8.

If this is really excreted by Water Shrew, though the number of this mammal is not clear, about one eighth of the liberated trout fries had taken by this during 8 days. This kind of predation was suspected since the beginning, but could not make sure by the research of the last year, probably because the amount of Zn^{65} was not enough to detect the radiation from the spots under huge rocks which has enough shielding effect against the radiation.

Though the monitoring efforts were intended to detect the radiation from the excrete of the bird, *Cinclus pallasii*, which was found to be a most powerful predator by the last years investigation and if it could be so it is very easy to detect by more sensitive survey instrument we used in this year, however, neither the radiation nor the Co^{60} pin was found about all the excrete we found in the area up to 4 km distant from the experiment stream.

About this rather curious phenomena, Mr. Tadakazu ISHIHARA of Kyoto University, who joined us for the ecological survey of this bird in this year, found a couple and two youngs of this bird having this stream of experiment as a part of their feeding territory, but could not see their nest in or near the experimental area. This situation is different from that of the last year, when the nest was found right at the liberation point and the couple was feeding their babies. Moreover, the size of the trout fries liberated to the stream was much larger in this year (8 g average) than in the last year (1.2 and 2.7 g average for each size group). Moreover, in the stream far upper than the experimented area, where more abundant excrete of this bird was found, there were plenty of much smaller sized trout fries born by natural spawning in this year. These difference may be the reason why the predation of this bird about the liberated trout fry was slight or none.

It should be noted here that this field research of 1960 was carried on by Mr. NOSE, Mr. SHIMIZU, Miss MATSUBARA, Mr. YOTSUMOTO and myself, with cooperation of Mr. ONODERA, Fresh-water Fisheries Research laboratory, Mr. SUZUKI, Kanagawa Prefecture Fisheries Laboratory, and members of Fishermans Cooperative Association of Tsukui, and helped greatly from ornithological stand point by Mr. ISHIHARA, Kyoto University.

Comparisons of the two Methods

The experiences obtained by doing Co^{60} pin method in 1960 tell the following differences from that of Zn^{65} method done in 1959 as the method to trace the predation.

(1) Co^{60} method required more effort in the recovering of the radiation sources, as it is necessary to take all these off from the field, though it was

not so large area. So it would need tremendous effort when this method would be applied to much larger area like large river. While Zn^{65} injection method needs no effort of recovery, if the amount of radio-isotope is well limited.

(2) Co^{60} method was found to have a big advantage to use it as a method to know about the kind of the predator, as it was proved to be so that the Water Shrew, which was not found by Zn^{65} injection method in 1959, was found by this method. On the contrary, as far as the quantitative survey of the predation is concerned, Zn^{65} method can go more precisely to know about the number of victim taken by various predators, as already stated. While Co^{60} pin method can bring just a rough idea, as the number of marked fish for the liberation should be limited by the difficulty of the recovery of the radiation sources.

(3) Co^{60} pin method has a fatal defect for the interpretation of the cause of existence of found Co^{60} pin, when it was found outside of fish body, except in case the ecology of the possible predators were well studied already. Because this pin is not staying in the body of the predator so long after the predation, this is usually found in the excrete of the predators. While, Zn^{65} injected into trout fry can remain long enough in the body of the predator, so it is accurate to identify if the predator can be caught for close examination.

Beside the purpose to trace the predator, Co^{60} pin method was found so effective as the method to trace the redistribution of the liberated fish by the experiment of 1960, as the sensitive survey instrument can locate the fish living under water from some distance clearly.

The author believes that the choice of the method should be carefully done in any case of practice of this type of research, considering the public health problem well, then there would be so many problems to be cleared out by applying this type of method in future.

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Appendix

Contents of Previous Issues of the Records of Oceanographic Works in Japan

I. New Series

1. Published by the Japanese National Commission for UNESCO

Volume 1, No. 1 March 1953

Currents of Kuroshio and its branches in the seas adjacent to Hachijo Island in relation with fisheries (Report I)	M. UDA..	1
On the Cauchy-Poisson waves caused by eruption of a submarine volcano.	S. UNOKI and M. NAKANO..	11
On the eddy diffusion of pumices ejected from Myojin reef in the Southern Sea of Japan.....	M. HANZAWA..	18
On the variation of the oceanic condition in the sea near the fixed point "Extra".	J. FUKUOKA and T. TSUKI..	23
On the circulations of current in the China Eastern Sea.	Z. YASUI and M. MATSUNO..	28
On the annual variation in oceanographical elements at a fixed point (39N, 153E) in Pacific Ocean.....	M. KOIZUMI..	36
A current meter using a phototube.....	T. NAN'NITI..	44
"Kuroshio" undersea observation chamber....	N. INOUE, T. SASAKI and R. OAKI..	52
On the effect of waves upon the vertical distribution of water temperatures. ..	T. ICHIYE..	63
On the hydrographical conditions of the Sea of Japan in Spring and Summer 1949.	I. YAMANAKA..	71
Preliminary report of the observations on ocean waves at the Hachijo Island. ..	K. HIDAKA, K. YOSHIDA and K. KAJIURA..	81
Studies on Minute elements in Sea-water.	M. ISHIBASHI..	88
A study on the property of the coastal water around Hachijo Island.	Y. MIYAKE, Y. SUGIURA and K. KAMEDA..	93
Notes on the ecology of a plankton copepod, <i>Labidocera pavo</i> Giesbrecht.	S. K. WADA..	100
Observations on some marine luminous organisms Hachijo Island, Japan.	Y. HANEDA..	103
On some new or noteworthy algae from Hachijo Island.	Y. YAMADA and S. SEGAWA..	109
Notes on the flying fishes of Hachijo Island, with nomenclatorial remarks of the flying fishes of the mainland of Japan and Hokkaido. I. "Tobi-uo" <i>Prognichthys agoo</i>	T. ABE..	115
Molluscan Fauna of Hachijo-jima (preliminary report).....	I. TAKI..	124
The pelagic copepods of the Izu region.	O. TANAKA..	126
Personal differences and degree of difficulty in age determination of six fish		

species.	Y. HIYAMA..	138
Catches of a Japanese trawler in Saurashtra waters.....	K. R. SRIVATSA..	149

Volume 1, No. 2 December 1953

The rising of the coastal sea water temperature in winter.....	Jiro FUKUOKA and Iwao TSUKI..	1
Annual variation of sea surface water temperature in the current region.	Kozo HISHIDA..	7
On the oceanographic and meteorological conditions at the fixed station, "Extra" in the North Pacific Ocean I.....	Masami KOIZUMI..	13
A study on the foaming of sea water on the mechanism of the decay of foam layer of sea water.....	Tomosaburo ABE..	18
On the fluctuation of the Kuroshio Current off the Shionomisaki from 1950 to 1952.....	Jotaro MASUZAWA and Toshisuke NAKAI..	25
On the hydrographical conditions around an island.	Takashi ICHIYE..	33
Determination of selenium in sea-water.	Masayoshi ISHIBASHI, Tsunenobu SHIGEMATSU and Yasuharu NAKAGAWA..	44
Copper and Zinc in Pacific Waters.....	Yoshimi MORITA..	49
The chlorinity determination of sea water by micro-analytical method.....	Katsuko SARUHASHI..	52
On the diatom plankton of the Antarctic Ocean in the summer from 1946 to 1952.	Ryuzo MARUMO..	55
A peculiar chiton from Hachijo-jima with special reference to its distribution in the Pacific.....	Isao TAKI..	60
Larviparous oysters from the tropical West Pacific.	Seiji WADA..	66
Preliminary report on the <i>Branchyuran crustacea</i> of Hachijo Island. ..	Tune SAKAI..	73
Distribution of the recent foraminifera from the adjacent seas of Japan I.	Yokichi TAKAYANAGI..	78
Ecological analysis of pelagic shoals I.....	Hiroshi MAEDA..	86
Studies on naturally occurring suspended organic matter in waters adjacent to Japan.....	Tokimi TSUJITA..	94
Absorption of fish scale caused by starvation.	Ryushi ICHIKAWA..	101
Under-water animal sounds around Japan.....	Yoshio HIYAMA..	105

Volume 2, No. 1 March 1955

A Study on the Foaming of Sea Water—On the Mechanism of the Decay of Foams and their Size Distribution in Foam Layer of Sea Water.....	Tomosaburo ABE..	1
Upwelling induced by a circular wind system. ..	Koji HIDAKA and Yoshio AKIBA..	7
On the Vertical Distribution of Density and Upwellings in a Cold water Mass.	Shigeo HIKOSAKA..	19
Decay of the Large Horizontal Eddies in the Ocean.....	Kenko IWATA..	23
Simple Method of Investigation of Tidal Current in Narrow Channel and its Application to Ki-i Suido, Japan.....	Kenko IWATA..	29
On the distribution of the radio activity in the sea around Bikini Atoll in June, 1954.	Yasuo MIYAKE, Yoshio SUGIURA and Kazuhisa KAMEDA..	34
On the Content of Rubidium and Cesium in Sea-Water.	Masayoshi ISHIBASHI and Tadashi HARA..	45
Generation of the Tsunami in Compressible Water (Part II) ...	Hisashi MIYOSHI..	49

On the Vertical Distribution of the Intensity of Illumination in the Water and Water Temperature of Lake Ashinoko.	
Tadayoshi SASAKI, Noboru OKAMI, Seiichi WATANABE and Gohachiro OSHIBA..	57
On the Antarctic Circumpolar Current.	Kenzo TAKANO.. 71
An Example of Thermal Convective Current.	Kenzo TAKANO.. 76
Sur la Dimension des Tourbillons Cellulaires dans les Expériences de Henri BÉNARD.	Kenzo TAKANO.. 80
Sur les Tourbillons Thermoconvectifs dans la Nappe Fluide Horizontale Chauffée par en dessous.	Kenzo TAKANO.. 87
Sur la Différence Critique de Température dans les Expériences de Douchan Avsec au Sujet des Courants Thermoconvectifs.	Kenzo TAKANO.. 93
Sur les Courants Thermoconvectifs dans la Nappe Fluide Horizontale Chauffée par en dessous.	Kenzo TAKANO.. 97
The Relation between Chlorinity and Silicate Concentration of Water observed in Some Estuaries. ..	Hironari MAKIMOTO, Hiroshi MAEDA and Shitoku ERA..106
Seasonal Difference in the Amount of the Inorganic Phosphorus in Blood Serum of Fishes.	Ryushi ICHIKAWA..113
A new species of copepod, <i>Scolecithricella vespertina</i> from the Sea of Japan.	Otohiko TANAKA..117
Preface to the Reports of the Oceanographical and Biological Survey of the Kuroshio.	Koji HIDAKA..121
Hydrographic Survey in the Equatorial Pacific Ocean, the South China Sea and the Formosa-Satsunan Kuroshio Region.	Makoto ISHINO..125
Preliminary Report on the Kuroshio in the Eastern Sea of Japan (Currents and Water Masses of the Kuroshio System III).	Jotaro MASUZAWA..132
On the Subtropical Convergence and the Currents in the North Western Pacific.	Michitaka UDA..141
A Note on Forecasting Ocean Waves caused by Typhoons.	Sanae UNOKI and Mashito NAKANO..151
Markings on a Tidal Flat in Uranouchi Bay, Shikoku.	Kotora HATAI and Hideo MII..162
Observations on the Tidal Flats in Uranouchi Bay, Kochi Prefecture, Shikoku, Japan.	Kotora HATAI and Hideo MII..168
On a new Pacific flying-fish, <i>Prognichthys Sealei</i> retaining five unbranched finrays above in the pectoral throughout life.	Tokiharu ABE..185
Further Notes on the Branchyuran Crustacea of the Hachijo Island..	Tsune SAKAI..193
Distribution of Mopalioid Chitons in the Pacific.	Isao TAKI..193
Marine Plankton Copepods from Hachijo Island, Izu Is., with Special Reference to their Vertical Distribution.	Shigeru MOTODA and Masateru ANRAKU..210

Volume 2, No. 2 October 1955

Why are the Thermocline and Sea Surface Horizontal at the Equator?	Koji HIDAKA.. 1
Coastal Upwelling off the California Coast.	Kozo YOSHIDA.. 8
Note on the convective circulation (1)	Kenzo TAKANO.. 21
Drift and Diffusion of Radiologically Contaminated Water in the Ocean.	Hisashi MIYOSHI, Sadakiyo HORI and Shozo YOSHIDA.. 30
On a Simple Method for Estimating the Current Velocity at the Equator (II)..	Mizuki TSUCHIYA.. 37
Researches on the Fluctuation of the North Pacific Circulation. 1. The Fluctua-	

tion of Oyashiwo Current in Relation to the Atmospheric Circulation and to the Distribution of the Dichothermal Waters in the North Pacific Ocean.	
.....Michitaka UDA..	43
On the Theorem of Helland-Hansen and Ekman and Some its Applications.	
.....Shoitiro HAYAMI, Hideo KAWAI and Masao OUCHI..	56
On a Problem Concerning the Vertical Circulation of Sea Water Produded by Winds with Special Reference to its Bearing on Submarine Geology and Submarine Topography.	
.....Mashito NAKANO..	68
On the Possible Origin of the Intermediate Water in the Kuroshio.	
.....Takashi ICHIYE..	82
On the rapid determination of the dynamic depth anomaly in the Kuroshio Area.	
.....Masashi YASUI..	90
Notes on the Cross-Current Structure of the Kuroshio (Currents and Water Masses of the Kuroshi System V)..	
Jotaro MASUZAWA and Toshisuke NAKAI..	96
Oceanic Condition in the North-Western Pacific Ocean (Preliminary Report)	
.....Jiro FUKUOKA..	102
The Radiochemical Analysis of Radio-Nuclides in Sea Water Collected Near Bikini Atoll.	
Yasuo MIYAKE and Yoshio SUGIURA..	108
On the First Larva of <i>Parribucus Ursus-Major</i> (Herbst)	
.....Hiroaki AIKAWA and Kiyoshi ISOBE..	113
Distribution of Plankton Diatoms in Relation to Hydrographic Conditions in the Sea Area East of Honshu in the Summer of 1954.....	
Ryuzo MARUMO..	115
On a Manganese Nodule and Perotroclus dredged from the Banks Near the Izu Islands, Japan.....	
Hiroshi NIINO..	120
Boric Acid in Bittern (I).....	
Masayoshi ISHIBASHI, Koichi EMI, Hiroyoshi KATAOKA and Seizo KITADANI..	127
Optical properties of the water in the Kuroshio Current.....	
Tadayoshi SAKAI, Noboru OKAMI, Seichi WATANABE and Gohachiro OSHIBA..	133

Nolume 2, No. 3 October 1955

Seasonal Variations of the Sea Level Along the Japanese Coasts.....	
.....Masamori MIYAZAKI..	1
A Method to Determine the Time of Ring-Formation in Hard Tissues of Fishes Especially of the Age Determination of Pacific Tunas.	
.....Yukio NOSE, Sayako TOMOMATSU, Koya MIURA and Yoshio HIYAMA..	9
Comparative Studies on the Red Tide Appearing in the Waters Adjacent to Western Japan.....	
Tokimi TSUJITA..	19
On Some Recent Fish Otoliths.....	
Kotora HATAI..	28
Studies on the Removing Scale from Sea Water or Natural Brine.	
.....Jinkichi YOSHIMURA..	35
On the Electrolysis of Dilute Solution Chloride Electrolysis by Diaphragm-Amalgam Type Process.	
Taizo KUME and Masaru OKADA..	40
Quantitative Analysis of KCl and NaCl Mixtures with the X-Ray Spectrophotometer.....	
Shiro TAKASHIMA, Yuji HAGINO and Genzo HASHIZUME..	45
Biological Studies on the Practical Utilities of Poisonous Invertebrates I-A Preliminary Note on the Toxical Substance Detected in the Trumpet Sea-Urchin, <i>Toxopneuster Pileolus</i>	
Katsuhiko OKADA..	49
Sedimentological Examination on the Influences of Oceanographic Environments Upon Marine Humus Distribution.....	
Kenji KATO..	53
On the Exploration of the Naruto Straits. ..	
Takeo GODA and Mamoru KUROIWA..	63

Preliminary Note on the Bottom Topography of the "Damo" Area.	
.....Kaoru TATARA..	71
Annual Variation of Sea-Surface Water Temperature and Air Temperature in the Western Parts of Pacific Ocean.	Kozo HISHIDA.. 79
Studies on the Marine Sulfate-Reducing Bacteria I. Distribution of Sulfate-Reducing Bacteria in the Coastal Region of the Sea. Masao KIMATA and Hazime KADOTA..	85
Studies on the Preparation of Sodium Alginate I. Treatment of the Raw Material Sea-weed.	
.....Taizo KUME and Hiroshi NISHIMADA..	94

2. Published by the Science Council of Japan

Volume 3, No. 1 March 1957

Preface.	Shinkishi HATAI
An Example of Cold Eddies South of the Kuroshio.	Jotaro MASUZAWA.. 1
On the Rapid Estimation of the Dynamic Topography in the Seas Adjacent to Japan.	Masashi YASUI.. 8
A Note on the Horizontal Eddy Viscosity in the Kuroshio.	Takashi ICHIYE.. 16
On the Storm Surges Caused by the Typhoon 5415 "Marie" (I)...	Takashi ICHIYE.. 26
On the Storm Surges Caused by the Typhoon 5415 "Marie" (II)...	Takashi ICHIYE.. 36
On the Tidal Mixing of Sea Water Through Narrow Channels....	Shoiti HAYAMI.. 49
Investigation of the Sea Floor of Tsugaru Straits by Under-water Camera.	
Tadayoshi SASAKI, Seiichi WATANABE, Gohachiro OSHIBA and Noboru OKAMI..	56
Maximum Permissible Concentration of Sr ⁹⁰ in Food and its Environments.	
.....Yoshio HIYAMA..	70
Up-take of Strontium by Marine Fish from the Environment (I).	
.....Yoshio HIYAMA and Ryushi ICHIKAWA..	78
Mechanism of Utilization of Plankton by Some Fishes.	
.....Fujio YASUDA and Yoshio HIYAMA..	85
Optical Properties of the Water in the Kuroshio Current (II).	
Tadayoshi SASAKI, Noboru OKAMI, Seiichi WATANABE and Gohachiro ŌSHIBA..	92

Volume 4, No. 1 December 1957

A Supplemetary Note on the Foaming of Sea Water.	Tomosaburo ABE.. 1
A Note on the Westward Intensification of Ocean Current.	Jiro FUKUOKA.. 7
Northward Flow in Lower Layers as an Indicator of Coastal Upwelling.	
.....Kozo YOSHIDA and Mizuki TSUCHIYA..	14
Observations along the Southern Coast of Izu Peninsula, Shizuoka Prefecture, Japan.	Kotora HATAI and Hideo MIH.. 23
Preliminary Observations along the Sea Coast of Wakayama Prefecture, Japan. Japan.	Kotora HATAI and Hideo MIH.. 30
Evidence for an Ancient Shoreline on a Marine Terrace, West Coast of Waka- yama Prefecture, Japan. ..	Kotora HATAI, Yushi FUNAYAMA and Hideo MIH.. 39
A Note on the Development of Certain Marine Pot-holes along the West Coast of Wakayama Prefecture, Japan.	
.....Kotora HATAI, Yasui FUNAYAMA and Hideo MIH..	45
A Measure on Level of Strontium-90 Concentration in Sea-Water Around Japan, at the End of 1956.	Yoshio HIYAMA and Ryushi ICHIKAWA.. 49

- The Amount of Up-take of Strontium by Marine Fish from the Environment of Various Concentration of Calcium and Strontium. Ryushi ICHIKAWA and Yoshio HIYAMA.. 55
- Morphometric Characteristics of the Albacore, *Germo germa*, from the North-west Pacific. Kenji KUROGANE and Yoshio HIYAMA.. 67

Volume 4, No. 2 March 1958

- Computation of the Wind Stresses over the Oceans. Koji HIDAHA.. 77
- On the Seasonal Variation of Surface Divergence of the Ocean Currents in Terms of Wind Stresses in the Oceans. Koji HIDAHA and Kiheiji OGAWA.. 124
- Depth Determination in Oceanographic Observations. Toshihiko TERAMOTO.. 170
- A Study on Upwelling. Kozo YOSHIDA.. 186
- Productibility of Plankton-feeders in the Off-shore Area of Japan. Yoshio HIYAMA and Fujio YASUDA.. 193
- Morphometric Comparison of the Alcacore from the Northwest, the Equatorial and the Southwest Pacific..... Kenji KUROGANE and Yoshio HIYAMA.. 200

Volume 5, No. 1 February 1959

- A Note on the Ice and Ocean Conditions in the Sea near Syowa Base in Antarctica. Jiro FUKUOKA.. 1
- Experiments on Evaporation from Wavy Water Surface. Setsuo OKUDA and Shōitirō HAYAMI.. 6
- A Hypothesis on Transmission of Energy of Tsunami Waves. Kozo YOSHIDA.. 14
- A Practical Method for a Determination of Reflection of Long Gravitational Waves. Kiheiji OGAWA and Kozo YOSHIDA.. 38
- An Estimation of the Average Compensation Depth by a Study of the Vertical Distribution of Dissolved Oxygen. Akira OKUBO.. 51
- Preliminary Report on the Shelf Sediments off Kamo Yamagata, Japan. Hideo KAGAMI.. 60
- Morphometric Comparison of the Alcacroe from the Indian and the Pacific Ocean. Kenji KUROGANE and Yoshio HIYAMA.. 68
- A Measure of Role of Utilization of Plankton-feeders by Carnivorous Fishes in a Certain Area..... Fujio YASUDA and Yoshio HIYAMA.. 85
- Measurements of Submarine Light Polarization. Tadayoshi SASAKI, Noboru OKAMI, Seiichi WATANABE and Gohachiro OSHIBA.. 91

Volume 5, No. 2 February 1960

- Angular Distribution of Scattered Light in Deep Sea Water. Tadayoshi SASAKI, Noboru OKAMI, Gohachiro OSHIBA and Seiichi WATANABE.. 1
- The Oceanic Waves of Days to Manths' Periods. Kozo YOSHIDA.. 11
- On the Distribution of Divergence and Convergence of Surface Drift Vectors in the Western Pacific Oceans. Hideo SUDO.. 25
- The Thickness of the Upper Homogeneous Layer. Shigeo MORIYASU.. 44
- Existence of a Contra Solem Vertical Motion in the Coral Sea.. Tadao TAKAHASHI.. 52
- Inorganic Constituents in Seaweeds.. Masayoshi ISHIBASHI and Toshio YAMAMOTO.. 55
- Determination of Manganese in Sea-Water. Masayoshi ISHIBASHI, Tsunenobu SHIGEMATSU and Yasuharu NISHIKAWA.. 63
- Determination of Arsenic in Sea-Water. Masayoshi ISHIBASHI,

Tsunenobu SHIGEMATSU, Yasuharu NISHIKAWA and Yoshihiro ISHIBASHI..	66
Studies of Vitamin B ₁₂ in Natural Water.....	
.....Ken-ichi KASHIWADA, Daiichi KAKIMOTO and Akio KANAZAWA..	71
Biology of <i>Sebastiscus Marmoratus</i> CUVIER et VALENCIENNES.	Shin-ichi MIO.. 77
Biology of <i>Sebastes Inermis</i> CUVIER et VALENCIENNES.....	Shin-ichi MIO.. 86
An Ider on the Maximum Permissible Concentration of Radio-Active Materials in Sea-Water.....	Yoshio HIYAMA.. 98
Morphometric Comparison of the Yellowfin Tuna from the Banda Sea and its Adjacent Waters.	Kenji KUROGANE..105
Strontium-Calcium Discrimination in the Rainbow Trout.	Ryushi ICHIKAWA..120
The Feeding Mechanism in Young Fishes.....	Fujio YASUDA..132
The Relationship of the Gill Structure and Food Habits of some Coastal Fishes in Japan.....	Fujio YASUDA..139
The Feeding Mechanism in Some Carnivorous Fishes.....	Fujio YASUDA..153
Notes on Some Edible Marine Fishes Collected Between the Bonin Islands and the Mouth of Sagami Bay. II.	Tokiharu ABE..161
On the Larva Possibly Referable to <i>Marlina Marlina</i> (JORDAN HILL).	
.....Shoji UYEYANAGI and Hiroshi YABE..	167

II. Special Number

Published by the Japanese National Commission for UNESCO

Special Number 1 March 1957

Results from the Oceanographic Observations in the North Pacific (1955-56) with Umitaka-maru and Shinyo-maru.	
.....Michitaka UDA, Yoshimi MORITA and Makoto ISHINO..	1
Kuroshio during 1955.	Daitaro SHOJI.. 21
On the exchange of the water masses between the Seto Inland Sea and the open seas through the Akashi Straight and the Naruto Straight.....	
.....Shoichiro HAYAMI, Y HUKUO and D YODA..	27
On the relationship between the plankton distribution and hydrographic condition in the adjacent seas of Japan.	Takashi ICHIYE.. 34
Measurements of the Angular Distribution of Daylight in the Sea.....	
.....Tadayoshi SASAKI, N. OKAMI, S. WATANABE and G. OSHIBA..	42
A Study on the Method of the Extraction of Potassium from Bittern as Perchlorate.....	Masayoshi ISHIBASHI and Fuji MORII.. 46
Studies on the Direct Manufacture of Sodium Hydroxide and Other Materials from sea Water. Part 3. Effects of Co-Existing Ions for Electrolysis.	
.....Taizo, KUME. Tokeo KUNO and Tamotsu TANAKA..	50
Studies on the Direct Manufacture of Sodium Hydroxide and Other Materials from Sea Water. Part 2. Electrolysis by Mercury Process Using a Diaphragm.	Taizo KUME and Masaru OKADA.. 53
Studies on the Direct Manufacture of Sodium Hydroxide from Sea Water.....	
.....Taizo KUME..	57
Sediments on three Submarine Banks as promising Fishery Grounds (Sediments on the Umitaka, Yoneyama and Senjyusho).	Hiroshi NIINO.. 58
Geology and Natural Resources along the Coast and Bays of Southern Izu	

Penninsula.	Kotora HATAI..	63
The Methods of Utilization of Plankton by Fishes.	Yoshio HIYAMA and Fujio YASUDA..	67
The life history of fishes in the areas affected by the warm currents (Kuroshio) in the sea on the South of Japan.	Keitaro UCHIDA..	71
Report on some groups of sea-weeds belonging to Rhodophyta (red algae) collected around Hachijojima and Mikurajima Islands and also off Tosa district....	Yukio YAMADA..	73
Report on the sea-weeds, especially corallinae found around the Izu-Miyake and Tosa-Kashiwa Islands.	Sokichi SEGAWA..	77
The Relationship between the Inland Sea of Seto and the Open Sea II.		
Deposits of Molluscan Shells in the Hiuchi-nada and the Bingo-nada.....	Denzaburo MIYADI and Tadashige HABA..	79
The Relationship between the Inland Sea and Open Sea III.	Denzaburo MIYADI and Tadashige HABA..	87
Studies on the fishes whose young are believed to be carried in large numbers by the "Kuro-shiwo".	Tokiharu ABE..	93
Studies on the luminous organisms found in waters adjacent to the Pacific coasts of Japan.....	Yata HANEDA..	97
An Outline of Biological Survey and Distribution of Wet Weight of Macroplankton in the "Kuroshio" Area, South of Honshu, May 1955.....	Zinjiro NAKAI..	103
Distribution of Copepoda in the "Kuroshio" Area, South of Honshu, May 1955..	Koji HONJO, Hankai OHTA, Takashi KIDACHI, Kiiti UMEDA and Seitoku KUDOH..	120
Distribution of Euphausiacea in the "Kuroshio" Area, South of Honshu, May 1955.	Koji HONJO..	130
Holoplanktonic Gastropoda in the "Kuroshio" Area, South of Honshu, May 1955.	Takashi OKUTANI..	134
Fish Larvae and Eggs in the "Kuroshio" Area, South of Honshu, May 1955....	Shigemasa HATTORI, Taisuke WATANABE and Takashi OKUTANI..	143
A Preliminary Report on the Biology Survey in the "Kuroshio" Area, South of Honshu, June-July 1955.	Zinjiro NAKAI, Shigemasa HATTORI, Koji HONJO, Taisuke WATANABE, Hankai OHTA, Kiiti UMEDA and Seitoku KUDOH..	159
A Preliminary Report on the Oceanographic Survey in the "Kuroshio" Area, South of Honshu, June-July 1955.	Nobuo WATANABE, Toshiyuki HIRANO, Ryonosuke FUKAI, Kozi MATSUMOTO and Fumiko SHIOKAWA..	197

Special Number 2 March 1958

A Study on the Productivity of the Tanabe Bay (Part I).	Denzaburo MIYADI and Shoitiro HAYAMI..	1
A Study on the Productivity of the Tanabe Bay (Part I). I. Oceanographic Conditions of the Tanabe Bay—Results of the Survey in the Autumn of 1956..	Shin-ichiro FUSE, Isamu YAMAZI and Eiji HARADA..	3
A Study on the Productivity of the Tanabe Bay (Part I). II. Tidal Currents in the Tanabe Bay.	Akikazu NAKAMURA..	10
A Study on the Productivity of the Tanabe Bay (Part I). III. On the exchange of Water and Productivity of a Bay with Special Reference to the Tanabe Bay.	Shoichiro HAYAMI, Yoshiaki FUKUO and Daisaku YODA..	15
A Study on the Productivity of the Tanabe Bay (Part I) IV. On some Plankton Indicating the Water Exchange.	Isamu YAMAZI..	25
A Study on the Productivity of the Tanabe Bay (Part I). V. Distribution and		

Abundance of Pocky Reef Fishes in the Tanabe Bay.....	
.....Ryonosuke OKUNO, Shin-ichiro FUSE and Eiji HARADA..	36
A Study on the Productivity of the Tanabe Bay (Part I). VI. Zonal Arrange- ment of Intertidal Benthic Animals in the Tanabe Bay. ..Tadashige HABA..	43
A Study on the Productivity of the Tanabe Bay (Part I). VII. Horizontal Dis- tribution of Intertidal Communities on Rocky Shores Around the Tanabe Bay (Preliminary Report). ..Fuzio UTINOMI..	50
The Ocean Waves and the Secondary Undulations of Tides on the Southern Coast of Kii Peninsula. Masito NAKANO and Sanae UNOKI..	57
A Note on the Hydrographic Conditions of Kuroshio and its Northern Boundary 1956. Michitaka UDA..	68
On the "Shiome" at the Boundary Zone of the Kuroshio and the Coastal Water off Shionomisaki.....D. SHOJI, Ryuzo WATANABE and K. HASUIKE..	78
Coastal Upwelling, Coastal Currents, and their Variations. Kozo YOSHIDA..	85
On the Decay of a Cyclonic Vortex in the Southern Sea of Japan (Hydrography of the Kuroshio in relation to its Productivity, II). Takashi ICHIYE..	90
Turbidity Distribution and its Relation to Some Oceanographical Factors in the Late Summer of 1956. Satoshi NISHIZAWA and Naoichi INOUE..	101
A Submersible Turbidimeter.....	
Tadayoshi SASAKI, Gohachiro OSHIBA, Seiichi WATANABE and Noboru OKAMI..	116
Spectral Energy Distribution of Submarine Daylight off Kii Peninsula.	
Tadayoshi SASAKI, Noboru OKAMI, Gohachiro OSHIBA and Seiichi WATANABE..	120
Fluctuations in the Fishery for Young Sardine off Shichiri-Mihama, Kii Penin- sula. Zinjiro NAKAI..	128
The Variation of Hydrographic Condition in Urado Bay, ...Hiroshi YAMAMOTO..	145
Studies on the Utilization of the Shallow-Water Deposits (Continued).	
Masayoshi ISHIBASHI, Shunzo UEDA, Yoshikazu YAMAMOTO and Fuji MORII..	153
Marine Algae of the Kii Peninsula, with Special Reference to the Corallinaceae (Preliminary Report).....Sokichi SEGAWA..	167
Studies on the Luminous found in Waters Adjacent to the Pacific Coasts of Japan, Report II. Yata HANEDA..	171
Two New Subspecies of Fishes from the Path of the "Kuro-shiwo".....	
.....Tokiharu ABE..	175
A Preliminary List of the Marine Algae from the Shores of the Kii Peninsula. Yukio YAMADA..	181
The Formation of Sulfide by Sulfate-Reducing Bacteria in the Estuarine Zone of the River Receiving a Large Quantity of Organic Drainage.....	
....Masao KIMATA, Hajime KADOTA, Yoshihiko HATA and Hideo MIYOSHI..	187
Notes on the Polychaetous Annelid of Kii Peninsula. Keizo TAKAHASHI..	200
Shelf Sediments of the Gulf of Kumano, Japan I.	
.....Noriyuki NASU and Yasukazu SAITO..	205
A Weathering and Marine Erosion Product Observed along the Coast of Mie Prefecture, Japan. Kotora HATAI and Hideo MII..	211
Tongallen from the Atsumi Peninsula, Aichi Prefecture, Japan.	
.....Kotora HATAI and Shozo HAYASAKA..	217
An Erosion Feature Observed in Wakayama and Shizuoka Prefectures, Japan. ..	
.....Kotora HATAI and Hideo MII..	221
Strontium and Calcium Distribution in Western Pacific, Indian and Antarctic Oceans. Ken SUGAWARA and Nobuko KAWASAKI..	227

Special Number 3 October 1959

- A Study on the Productivity on Tanabe Bay (Part II).
 I. Tidal Currents in Tanabe Bay (2nd Report).....Akikazu NAKAMURA.. 1
- A Study on the Productivity of Tanabe Bay (Part II).
 II. On Some Plankton Indicating the Water Exchange in Tanabe Bay in August, 1957. Isamu YAMAZI.. 23
- A Study on the Productivity of Tanabe Bay (Part II).
 III. Oceanographic Conditions of Tanabe Bay (2) Stratification and Fluctuation of Hydrological Conditions on Two Sectional Survey Lines..... Shin-ichiro FUSE.. 31
- A Study on the Productivity of Tanabe Bay (Part II).
 IV. On the Voluminal Change of Plankton Communities Caused by Tidal Current. Isamu YAMAZI.. 47
- A Study on the Productivity of Tanabe Bay (Part II).
 V. Occurrence of the First Stage Phyllosoma Larva of *Panulirus Japonicus* (von SIEBOLD) in Tanabe Bay. Eiji HARADA.. 57
- A Study on the Productivity of Tanabe Bay (Part II).
 VI. On the Exchange of Water and the Productivity of a Bay with Special Reference to Tanabe Bay (2).....Shoiti HAYAMI and Yoshiaki FUKUO.. 61
- Pelecypod Shell Remains of Ariake Bay. Tadashige HABA.. 69
- Standing Oscillation of Bay Water Induced by Currents. Tomosaburo ABE.. 75
- Water Temperature Fluctuations of Muroto, Shikoku. Shigeo HIKOSAKA and Ryoji HIGANO.. 97
- Basic Studies for the New Method of Separation of Iodine from Natural Waters
 Absorption of Iodine to Polyvinyl Glycerin Acetal. Masayoshi ISHIBASHI, Koichi EMI, Tamotsu MATSUMOTO, and Takashi TERAMOTO..103
- Chemical Studies on the Seaweeds, III. The Content of Ash, Sodium and Potassium in Seaweeds. Masayoshi ISHIBASHI and Toshio YAMAMOTO..109
- The Hydrographic Condition and Variation of Calcium in Susaki Bay. Hiroshi YAMAMOTO and Yoshihiko IMAI..117
- Studies on the Utilization of the Shallow-water Deposits (Continued).
 On the Zinc Content of the Shallow-water Deposits. Masayoshi ISHIBASHI, Shunzo UEDA and Yoshikazu YAMAMOTO..123
- Studies on the Inorganic Chemical Constituents of Sea Fishes (Part I). Noboru IMANISHI..135
- Studies of Beach-Sands as Chemical Resources. On Sand-Irons in Beach-Sands along the Coast-line of Tottori Prefecture. Sakujiro YAMAMOTO..141
- Studies on Manufacture of Magnesium Hydroxide from Calcium Carbide and Sea Water or Bitter Brine and its Utilization (Part III).....Shiro TAKASHIMA..145
- Studies on the Direct Manufacture of Sodium Hydroxide and Other Materials from Sea Water (Part V). Influence of Co-existing Mg^{++} and Ca^{++} Ions on the Current Efficiency. Taizo KUME and Takeo HISANO..151
- Preparation of Potassium Chlorate by Electrolytic Oxidation from Bittern or Carnallite. Toshiharu MURAKAMI..155
- On the Natural Radioactivity of the Marine Sediments (Part I. Preliminary Report). Hiroshi NIINO..163
- Coastal Geology of the Kii Peninsula. Kotora HATAI..165
- Growth of Sulphate-Reducing Bacteria in Relation to the Oxidation-Reduction Potential of Culture Media. Masao KIMATA, Hajime KADOTA, Yoshihiko HATA and Hideo MIYOSHI..167
- Two New Types of Luminous Fish Organ, Notes on Biochemical Research.

Conducted, and a Clear Demonstration of Luciferin Luciferase Reaction Therein.	Yata HANEDA..	175
Studies on the Fish Community from the Viewpoint of Diet in the Eastern Part of the Inland Sea of Japan.	Tasuku HANAOKA, Tomoo HAYASHI, Yutaka MURAKAMI, Yoshiya TAKAHASHI and Yoshiaki YAMAGUCHI..	179
Biology of the Japanese Anchovy, <i>Engraulis Japonica</i> (HOULTUYN) in the Nankai Region (I).	Koji TAKESHITA and Hiroaki AIKAWA..	201
The Marine Algae of the Kii Peninsula.	Yukio YAMADA..	217
Studies on the Corallinaceous algae in the Warmer Seas around Japan.	Sokichi SEGAWA..	221
On the Presence of at least two species of Cubiceps (Nomeidae, Pisces) in the Path of the "Kuro-shiwo".	Tokiharu ABE..	225
Turbidity Distribution in the Subarctic Water of the North Pacific in the Summer of 1957.	Satoshi NISHIZAWA, Naoichi INOUE and Yoshio AKIBA..	231

Special Number 4 March 1960

A Study on the Productivity of Tanabe Bay (Part III) Tidal Currents in Tanabe Bay (3rd Report).	Akikazu NAKAMURA..	1
A Study on the Productivity of Tanabe Bay (Part III). Oceanographic Conditions of Tanabe Bay (3). Result of the Survey in the Summer of 1958.	Shin-ichiro FUSE and Eiji HARADA..	13
On the Exchange of Water and the Productivity of a Bay with Special Reference to Tanabe Bay (III).	Yoshiaki FUKUO..	29
Pelecypod Shell Remains in Tanabe Bay, Wakayama Prefecture..	Tadashige HABA..	39
Distribution of Sulphate-Reducing Bacteria in Maizuru Bay.	Masao KIMATA, Hajime KADOTA, Yoshihiko HATA and Hideo MIYOSHI..	53
On the Short Period Variations of Oceanographical Conditions in the Kuroshio Region (Preliminary Report).	Yoshitada TAKENOUTI and Masasi YASUI..	55
Researches on the Currents of Kuroshio.	Michitaka UDA and Makoto ISHINO..	59
Chemical Studies on the Ocean (Part 78). Chemical Studies on the Seaweeds (VI). The Content of Calcium, Magnesium and Phosphorus in Seaweeds.	Masayoshi ISHIBASHI and Toshio YAMAMOTO..	73
Chemical Studies on the Ocean (Part 79). Chemical Studies on the Seaweeds (VII). Iron Content in Seaweeds.	Masayoshi ISHIBASHI and Toshio YAMAMOTO..	79
Chemical Studies on the Ocean (Part 82). Chemical Studies on the Seaweeds Quantitative Determination of Beryllium in Seaweeds.	Masayoshi ISHIBASHI, Taitiro FUJINAGA and Tooru KUWAMOTO..	87
Chemical Studies on the Ocean (Part 83). Chemical Studies on the Seaweeds Quantitative Determination of Strontium in Seaweeds.	Masayoshi ISHIBASHI, Taitiro FUJINAGA, Tooru KUWAMOTO and Takahiko KUMAMARU..	91
Basic Studies on the Analysis and the Separation of Boric Acid (2nd Report). On the Complex Precipitate of Barium Boro Manno Saccharate and Barium Boro Aravotrihydroxyglutarate.	Masayoshi ISHIBASHI, Koichi EMI, Takeshi KUSAKA and Mitsuyuki MITOOKA..	95
Studies on the Utilization of the Shallow-water Deposits (Continued). On the Lead Content of the Shallow-water Deposits.	Masayoshi ISHIBASHI, Shunzo UEDA and Yoshikazu YAMAMOTO..	111
Distribution of Boron in Urado Bay. Especially about the Variation of Boron Content Accompanied by the Dilution of the Sea Water.		

.....Hiroshi YAMAMOTO and Yoshihiko IMAI..123	
Studies on Inorganic Chemical Constituents in Sea Fishes. The Chemical Elements, and the Relative Quantities of Calcium and Phosphorus in <i>Katsuwonus phlamis</i> (Part 4). Noboru IMANISHI..129	
Studies on the Direct Manufacture of Sodium Hydroxide and Other Materials from Sea Water (Part VI). Effects of Magnesium and Calcium Ions (Supplement) Taizo KUME, Takeo HISANO and Hideaki IZAWA..135	
Studies on Manufacture of Magnesium Hydroxide from Calcium Carbide and Sea Water or Bitter Brine, and its Utilization (IV). On Determination of pH of Reaction Medium and Effect of Size of CaC_2 Shiro TAKASHIMA..139	
Notes on Fishes from the Path of the "Kuroshio", with Special Reference to the Adaption or Preference of Some Flying-fishes for Cool Water. Tokiharu ABE..147	
Preliminary Report on Diurnal Feeding Activities of Genus <i>Saurida</i> Tomo'o HAYASHI, Yoshiaki YAMAGUCHI and Tasuku HANAOKA..151	
Biology of Japanese Anchovy, <i>Engraulis Japonica</i> (HOULTYIN), in the Nankai Region (II).....Koji TAKESHITA and Hiroaki AIKAWA..159	
Study on the Natural Radioactivity of Bottom Deposits (II). Hiroshi NIINO..165	
Evidence for Recent Changes in Sea-level of the Atsumi Peninsula, Atsumi-gun, Aichi Prefecture.....Katora HATAI, Shozo HAYASAKA and Takehiko IWAI..171	
Ground Water and Related Phenomena Observed along the Southern Coast of the Atsumi Peninsula, Aichi Prefecture. Katora, HATAI, Shozo HAYASAKA and Takehiko IWAI..179	
Some Cylindrical and Tabular Structures from Wakayama Prefecture, Kii Peninsula.....Katora HATAI and Takehiko IWAI..187	
Measurements of Perpendicular and Horizontal Angular Distributions of Submarine Daylight by means of a New Remote Control Instrument. Tadayoshi SASAKI, Seiichi WATANABE, Gohachirô OSHIBA and Noboru OKAMI..197	
Capped Cylinder Bucket. Magoshichi SATO..207	

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